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# VALLEY DEPOSITS IMMEDIATELY EAST OF THE CHANNELED SCABLAND OF WASHINGTON. I

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## ABSTRACT

Forty-one valleys whose drainage enters the eastern margin of scabland spillways are known to contain deposits made by glacial waters. Thirty-nine of these valleys never have carried drainage from glaciated areas, and the glacial waters in them must have been backed up from the scabland. These deposits possess such extraordinary features and relations that they, like the scabland topographic forms, are considered unique.

1. Each separate area of backwater is recorded by a widespread mantle of silt containing abundant grains of unweathered basalt (the country rock) and pebbles of foreign rock.

2. In each, this mantle extends up to a definite upper limit on the valley slopes and along the valley lengths.

3. These upper limits in each case agree closely with the upper limit of scabland where the valley enters.

4. The altitudes of the upper limits, constant throughout any one backwater area, vary with different pondings, forming a descending series from north to south along the scabland gradient.

5. Remarkable large mounded gravel deposits in the tributary valley debouchures possess foreset strata which dip *out* of the scabland and *up* the valleys. The composition, topography, topographical relations, structure, and size of these deposits are inexplicable without great *reverse currents*.

6. In no case could these reverse currents continue through and escape from the valleys they entered. They are simply a record of the rapid backflow into these valleys.

It is concluded that these glacio-aqueous deposits cannot be explained by conditions associated with ordinary glacial ablation or by a sequence of several Pleistocene epochs. They require the volume and rapid rise of glacial rivers across the plateau, which the writer has previously read from the scabland itself.

## INTRODUCTION

About forty different stream valleys draining westward into the easternmost of the channeled scabland tracts of the Columbia Plateau in Washington are considered in this paper. Many of them are

tributary to others, so that the total number of separate entrances into the scabland is twelve. The northernmost and southernmost of these entrances, measured along the scabland margin, are about 125 miles apart.

The deposits studied are thickest and coarsest close to the entrance of these valleys into the scabland. They extend up on the valley slopes as high as the loessial scarp bases along the adjacent scabland. They also extend back up these valleys approximately to the crossing of the contour which marks both their upper limit and the scarp bases at debouchure into the scabland. In all but three of these valleys, essentially the full length was examined, and the absence of these deposits farther upstream than the contour crossing was established for all but two. But the distance upstream to such crossings and the altitude of these crossings vary in different valleys.

Although the deposits are prevailingly coarser nearer the scabland, there are great variations in character and structure in any one portion of a valley. The diversity much exceeds anything ever previously reported for stream-valley deposits, so far as the writer is aware. A classification might be attempted on the basis of textural gradations; it might also be constructed on a basis of sedimentary structures. Either method may give the impression that these sediments are much like other silt, sand, gravel, and boulder accumulations, or that the genesis of their bedding structures is similar. The only terms available for description of these deposits have genetic associations, and it will be difficult to avoid an erroneous mental picture despite caution on the part of both writer and reader. The printed page cannot convey the picture which field acquaintance gives, for there probably is nothing in the reader's experience sufficiently resembling the structural, textural, and topographic features of these deposits. All gradations and interminglings possible exist. One must not think of the textural types as distinctly separated from each other. To do so is to miss one of the most significant facts regarding them.

The region will be treated areally, and the distribution and relations of the deposits will be considered valley by valley, beginning in the north and progressing southward along the eastern margin of the scabland. The area involved is shown in Figure 1. Three of the

twelve separate valley entrances into the scabland are major streams. Palouse River has one tributary sharing in the record, Snake River has eleven with three subtributaries, and Walla Walla River has four tributaries and seven subtributaries which carry the deposits. In addition there are eight separate minor valley entrances, and among

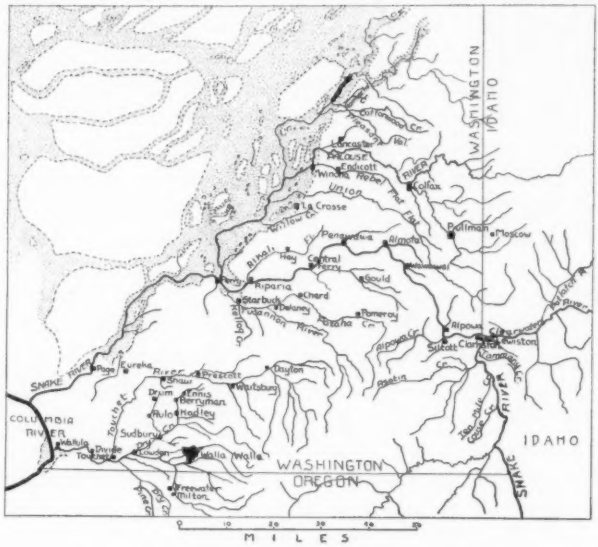


FIG. 1.—Valleys immediately east of channeled scabland

them four tributary valleys to be considered. This total of forty-one distinct valleys is only the total of those examined. Because of limitation of time, at least a dozen more tributaries and subtributaries of Snake River, joining below the known upper limits of loessial scarps and valley deposits, had to be left for later examination.

## PALOUSE RIVER DRAINAGE

### KAMIACHE CREEK VALLEY

The northernmost valley involved in this study is that of Kamiache Creek which enters the scabland just south of Rock Lake in northwestern Whitman County. The valley is only 7 miles long, though the stream flows 2 miles beyond the valley mouth across

scabland before it joins Rock Creek at the foot of the deep rock-basined lake. The contrast between the 7 miles of normal stream valley and the 2 miles of scabland can hardly be appreciated by one who has not seen it. Probably nothing like this has ever been described. Above the entrance to the scabland the valley is wide and flat bottomed, with alluvial deposits and low terraces on the floor, and with stream-etched slopes descending to it from the divides surrounding it. Through this valley the creek flows with a gradient normal to such minor streams in a mature topography. But out on the scabland there is no valley at all. Bare and rugged rocky buttes are distributed very irregularly among the rock basins and rock-bottomed spillways which separate them, making the anastomosing pattern which is typical of scabland. The extreme youthfulness of slopes here is emphasized by the cliffs, the blocky talus, and the unfilled basins. The utter unlikeness of this anastomosing pattern to stream valleys has been repeatedly pointed out by the writer as indicating that scabland is stream-channel topography. The impossibility of these channels recording a succession of routes has also been shown. They are contemporaneous and constitute simply the highly diversified rock floor of a very great glacial river whose surface was higher than the tops of the buttes.

Another topographic feature unique in the scabland system is the lines of steep loessial scarps which truncate the mature topographic forms to the east and overlook the scabland to the west. The base of the scarp immediately north of Kamiache Creek entrance is 1,880 feet A.T. Scabland descends thence down to Rock Lake, 145 feet lower, and in the rock basin of the lake to a reported depth of 250 feet. The total vertical range, therefore, is nearly 400 feet. The loessial scarps belong to the scabland forms, not to the older slopes along the creek valley.

If, then, Kamiache Creek above the scabland is much older than the scabland itself[1] and its bordering scarps, and if the volume and

[1] Kirk Bryan, in *U.S. Geological Survey Bull. 790-B*, clearly recognizes the existence of two kinds of drainage ways on the Columbia Plateau in Washington, one group (scabland) containing those "that were deepened and modified by the great streams diverted across the plateau during glacial time" and the other group, those "developed by normal erosion before the diversion of the glacial waters." He notes that some normal valleys "formed in a cycle of erosion that antedates the great coulees produced in the diversion of the glacial waters" are tributary to scabland spillways.

depth of the glacial rivers were as great as the writer has argued, the lower part of this creek valley must have been flooded with glacial water back from its entrance into scabland. Whatever may be expected or interpreted as a record of this flooding must not occur higher on the valley sides or higher upstream than approximately 1,880 feet, and it must be well marked below that altitude. These restrictions, set forth for Kamiache Creek valley, are applicable equally and are equally required for the forty other valleys considered.

Kamiache Creek valley is remarkably constricted at its debouchure into scabland where the stream crosses the line of loessial scarps. The bottom width in the valley itself is about 1,000 feet for 3 or 4 miles above the mouth, but in the mouth there is no bottom width. Instead, the stream enters a narrow gulch about half a mile long and 50–100 feet deep, through which it descends 100 feet in passing from the normal valley to the scabland. The south wall and the bottom of this gulch are of basalt; the north wall is of gravel.

If one climbs this northern gravel wall of the gulch, he finds himself on a remarkable spur which projects southward completely across the valley (save for the gulch itself) from the northern valley slope. This spur is 1,000 feet long (across the valley mouth) and about half a mile wide (measured with the stream length). Its summit is 150 feet higher than the scabland immediately west and 50 feet higher than the valley floor immediately east. This summit is irregularly rolling and has undrained depressions on it. On the east it descends steeply to the flat valley floor.

The spur which thus transforms Kamiache Valley at the mouth into a narrow gulch is composed of very angular and poorly sorted basalt pebbles, cobbles, and boulders. A pit on the west side of the spur near the base shows only foreset beds, dipping 21°–23° southward across the course of the valley, along the scabland length, and along the direction of the aligned loessial scarps. The top of this singular deposit is 1,865 feet A.T., 15 feet lower than the scarp base about half a mile to the northward. Between them are bare, or nearly bare, knobs and ledges of the highest scabland.

The flat floor and gentle slopes of Kamiache Valley afford only one adequate section in the first 4 miles east of this abnormal gravel deposit. The material exposed is sandy silt, poorly stratified and

poorly sorted.[1] Unweathered pebbles scattered irregularly through this material are composed of various rocks which certainly do not outcrop in the creek's drainage area. A small unweathered granite boulder was found at 1,850 feet A.T. East of this point, only alluvium of reworked loess, containing some basalt gravel, was found on the valley floor, and only loess on the valley slopes. There is no gravel deposit between the basalt and the loess and, therefore, no possibility of deriving the fresh granite, quartzite, etc., from the creek's drainage area. There is no possibility whatever that the great deposit of angular basalt gravel, containing boulders 3 feet in diameter, rising abruptly 50 feet above the flood plain, containing long deltaic foresets dipping southward across the valley, and clearly once completely blocking the valley, has been made by the creek. Just above the summit altitude of the gravel deposit are loessial scarps facing scabland to the west. This gravel mound, lying along the margin of the scabland and extending throughout almost its whole exposed vertical range, is a unit deposit with a definite topographic form modified subsequently only by erosion of the narrow gulch. It is a torrential deposit of some sort, built originally from the north completely across the mouth of the valley, and definitely related to the genesis of the scabland. The silt and erratic material up the valley do not occur higher than the summit level of this barrier.

[1] This silt was shaken through a series of ten sieves ranging from 1 mm. opening down to 0.074 mm. Material stopping on the 1 mm. sieve yielded recognizable particles of granite, quartzite, schist, mica, feldspar, clear quartz, and basalt. The quartz is abundant, its grains mostly sharp angular pieces with hackly surfaces, though some grains are rounded. Grains partly of quartz and partly of feldspar, and others of quartz with mica only partially embedded in it, tell of mechanical disintegration without weathering. The mica occurs in thick pieces, essentially broken mica crystals, instead of mica flakes. The percentage of basalt, the country rock, is low. It occurs as angular broken particles of fresh black rock and also as crumblike bits of reddish weathered rock. Material stopping on the 0.71 mm. sieve has a higher percentage of fresh broken bits of basalt, though down through the finer sieves the proportionate amount of basalt decreases rapidly while the proportion of quartz increases to 75 per cent on the 0.147 mm. sieve. Many of the particles of quartz on this sieve are shaped like splinters, two or three times as long as wide or thick. More than 50 per cent of the silt stops on or passes the 0.074 mm. sieve (200 meshes to the inch). In texture and in color it is identical with the loess of the nearby higher slopes.

COTTONWOOD CREEK VALLEY

This much larger stream enters the scabland 1½ miles southwest of the mouth of Kamiache Valley. The same striking contrasts between the topography upstream and downstream from its entrance exist as in Kamiache Creek. The interpretation of scabland as channel topography demands the same corroborative evidence in this valley. It is here but differently expressed.

The valley-mouth bar depending from the north and constricting the valley is much less definitely shown. It apparently never completely crossed Cottonwood Valley. There is no gulch across it and no steepened gradient comparable to the Kamiache features. But there is another bar, just south of the mouth of Cottonwood Creek valley, and its relations to scabland on the west and to loessial scarps on the east are clearly shown. Its upper surface falls but little short of the scarp base, its slopes are broadly convex and it partially buries scabland knobs. Bar deposits, therefore, testify to the depth of water in the glacial river at the mouth of Cottonwood Creek and indicate the necessity for silt deposits back in Cottonwood Valley pond.

Such silt is exposed in many places in the mature valley, extending back at least 6 miles from the bar. Since the valley-mouth bar does not appear ever to have constituted a dam, the silt can hardly be interpreted as deposited in a local pond. That it is not wholly from local drainage is abundantly shown by the presence of many pebbles of varied foreign materials.

The silt occurs in definite terraces near the junction of Pleasant Valley with Cottonwood Creek. Both streams have re-excavated their valleys to a depth of 30 or 40 feet in this silt, and many minor drainage furrows have been eroded across the terraces. Two miles or so farther up each valley, the silt deposit constitutes a fairly complete floor, only slightly trenched by streams.

A 30-foot clean section of the silt shows distinct stratification. Some strata are composed of the local loess, rehandled by water and largely free from other material. Other strata have abundant grains and small pebbles of basalt scattered through the dominant loess. Much less numerous are the angular grains and pebbles of foreign

rock. They are rarely grouped together in laminae, lenses, or pockets. The undisturbed stratification shows that the widespread dissemination of coarser material through the loess-silt is a result of original transportation and deposition together.[1]

This association of grains and small pebbles of foreign rock and of both fresh and weathered basalt in the reworked loess is consistently present in all the silt deposits described in this paper. Whatever the cause of ponding in these valleys, it did not allow the degree of sorting and segregation that occurs in essentially all water-laid deposits. Glaciers and mud flows leave unsorted valley deposits but obviously did not make these.

The silts in Pleasant Valley were identified within 2 miles of St. John at an upper limit of 1,940 feet A.T. None was found higher on the slope or higher up the valley, a feature indicating transportation up-valley rather than down-valley.[2]

PALOUSE RIVER VALLEY

This stream enters the scabland about 8 miles south of the entrance of Cottonwood Valley and at an altitude approximately 200

[1] Three samples of the silt in Cottonwood and Pleasant valleys were sieved, as described for the silt of Kamiache Valley. One sample came from the highest of the deposit, 1,940 feet A.T., and in all respects was like the other two. Material stopped on the 1 mm. screen was about 50 per cent basalt, all in small pieces. The larger fragments, ranging up to 1 inch in diameter, consist of granite, quartzite of several colors and textures, vein quartz, phyllite, muscovite schist, broken feldspar, and muscovite crystals. One mica fragment measured $\frac{1}{4} \times \frac{7}{16}$ inch. Part of the basalt present is in unweathered or but slightly weathered angular pieces whose surfaces are cleanly broken, part consists of much weathered pellets and crumbling pieces. In material stopped on the 0.71 mm. sieve and examined under a magnification of ten, some of the weathered pieces break when compressed in small tweezers or yield a claylike material when scraped. Yet some of the clean-fractured basalt particles are so hard that their sharp edges will take a steel mark from a knife blade. Sharply angular quartz grains are found in all the coarser sieves, though such sizes are unknown in the loess above the glacial water level, as are the bits of foreign rock. The material stopped by, and passing, the 0.074 mm. sieve amounts to 58 per cent, 60 per cent, and 85 per cent in the three samples.

[2] This altitude of 1,940 feet is 60 feet above the highest scarp bases in this latitude. If there is no error in measuring, it seems to indicate that the glacial waters in the scablands reached well up on some of the scarps and that scarp bases, therefore, are not the upper limits of the glacial rivers in the scablands. More evidence indicating this is found in a minor valley entering the scabland about 2 miles south of the entrance of Cottonwood. The scarp bases here are 1,865 feet A.T., and the silt occurs for half a mile back in it and 20 or 30 feet higher. It is the same type of deposit, carrying abundant coarse angular grains of black basalt with a few granite and quartzite particles.

feet lower. The profile of the floor of Palouse Canyon continues out into the scabland without interruption; the canyon walls, though scoured into scabland forms, are traceable for some miles out into the glacial drainage way; and the normal canyon for 1½ miles above the scabland is clogged with thick gravel deposits. Though the gravel is now dissected, there are extensive remnants of the original surface, and these consistently descend toward the east, or up-canyon. The bar profiles are unmistakable. A large remnant on the south side of the canyon also slopes down toward this wall, creating a broad fosse between the gravel mound and the north-facing cliffs. The maximum thickness of this gravel is more than 75 feet. Sections are few and unsatisfactory. One exposure in the eastern and lower part of the deposit shows irregularly bedded basaltic gravel, coarse basaltic sand, and fine brown sand and silt, with a total thickness of 40 feet. Cobbles and pebbles of various kinds of foreign rock are present.

There is no road or railroad along Palouse Canyon for nearly 20 miles, and no examination was made for silts up-valley from the significant gravel deposits. But Lancaster Creek, a tributary entering from the northeast, contains this silt for at least 4 miles above its junction with the Palouse, or 5 miles by valley from the scabland margin. A section 15 feet thick at 1,615 feet A.T. showed disseminated coarse grains and pebbles of fresh basalt, and a few foreign fragments.[1]

### REBEL FLAT CREEK VALLEY

The preglacial course of Palouse Canyon makes a right-angled turn to the south less than a mile out in the scabland and another right angled turn back toward the west about 5 miles farther downstream. At this latter angle, where the town of Winona stands, Rebel Flat Creek enters Palouse River from the east. Scarp bases just south of Winona, marking high-level divergences southward

[1] Seventeen per cent by weight of the total sample of this silt stopped on the 1 mm. sieve. Half of this consists of 6 angular pebbles of foreign rock. Seventy-five per cent of the remainder is basalt. By the mechanical tests outlined in describing silts of Cottonwood and Pleasant valleys, most of this basalt is but slightly weathered, though some is friable and soft. The little weathered basalt looks perfectly fresh in color; the crumbling particles are dark red in color and commonly do not show the clean-broken chiplike shapes with sharp edges. Seventy-one per cent of the total sample is essentially loess in texture and color.

across the preglacial divide toward Union Flat Creek, are 1,605 feet A.T. The floor of Rebel Flat is adjusted to the Palouse Valley floor at Winona. Its tributary valleys and all its slopes show that it is but a part of the general maturity east of the scabland, far older than the youthful scabland. Silt deposits for a few miles east of Winona show that, like the valleys thus far described, Rebel Flat was here in essentially present proportions before the glacial discharge eroded the scablands.

The silt deposits extend up this valley for at least 3 miles east of the scabland. One section, a mile east of Winona, shows 15 feet of material, well stratified and well sorted, mostly rehandled loess. Near the bottom of the section are two or three layers containing coarse black basaltic sand. A very few small erratic pebbles were found.

Near the up-valley limit of the silt are dissected terraces, 25–30 feet high, cuts in which show poorly stratified and poorly sorted pebbly silt. No deposits and no terraces of this character were found farther east. The highest altitude in which the silt is known to occur in this valley is 2 miles west of Endicott, 1,500 feet A.T. and 105 feet below the scarp bases at Winona.[1] But shallow road cuts are the only sections available, and in these the overlying loess or dust generally is not cut through. No gravel deposit has been identified back in the mouth of Rebel Flat Valley.

### UNION FLAT CREEK VALLEY

The lower 15–18 miles of Rebel Flat and Union Flat are nearly parallel, with an upland about 5 miles wide separating them. The southern valley is much longer, however, and its gradient near scabland is lower than that of Rebel Flat. Correspondingly, silts recording glacial backwater extend farther east in Union Flat. They are exposed in road cuts along approximately 14 miles of this valley,

[1] The material stopped by the 1 mm. sieve, 8 per cent of the total sample, is dominantly of basalt, most of which is but little weathered, though a few particles are much decayed. Foreign rock is represented by bits of granite and quartzite. Ten per cent of this coarsest material is quartz. On the 0.71 mm. sieve a larger percentage of the basalt is unweathered, and 35 per cent of the total is composed of quartz grains, most of them nearly 100 per cent angular. The material stopped by the 0.5 mm., 0.35 mm., 0.25 mm. and 0.177 mm. sieves is 50 per cent quartz grains. Sixty-seven per cent of the total sample is reworked loess.

highest at the easternmost exposure, where they are 1,680 feet A.T.[1] Foreign pebbles are abundant in the silt at 1,500 feet A.T., about 8 miles above the mouth.[2]

No localized gravel deposit was seen at the mouth; but the valley is a mile wide, and the glacial river channel it enters here is wide and shallow. The conditions favoring a marked valley-mouth bar were not present.

### WILLOW CREEK VALLEY

Glacial waters crossed the divide between lower Union Flat Creek and the minor subparallel Willow Creek valley on the south, entering the latter in no less than five places. More than half the length of this valley was traversed by glacial streams, only the upper 6 miles of its course escaping the great changes wrought by the invading flood from the north. Yet the volume which followed the Palouse, swinging around the western end of the original divide, was so much greater that a bar was built in the mouth of Willow Creek valley close to its junction with Palouse River. It is a larger and even more striking feature than that in Kamiache Creek valley.

This strictly local great mound of gravel has a total vertical range of 200 feet. It originally completely blocked Willow Creek valley, and out in the middle of this valley was 100 feet thick. The bar is nearly a mile wide (along the valley length). The length of the bar lies across Willow Creek valley and extends somewhat up-valley on the south side. At the north end the bar form is lost on the upper slopes, where the deposit is a thick mantle of gravel that rises north-

[1] The coarser constituents of this sample are almost entirely of basalt, and almost all of this basalt is weathered. Yet coarse, angular quartz particles and one particle of quartzite are contributions from glacial water backed up this valley. Some of the basalt is but slightly weathered. Less than 4 per cent of the total sample was stopped on the first two sieves; 70 per cent passed all sieves; and 12 per cent passed all but the finest sieve (0.074 mm.). This fine material, as in previous sievings, is essentially loess. The altitude here is 140 feet higher than the scarp bases on minor loessial islands in scabland about a mile below the entrance of Union Flat Creek, and 75 feet higher than scarp bases at Winona, 5 miles north of the entrance. This is further evidence that the surfaces of the glacial rivers were somewhat higher than the bases of the loessial scarps. Though this makes more difficult any precise reconstruction of the surface gradients of these scabland streams, it strengthens the case for great volume.

[2] Fifteen per cent of this sample is stopped on the 1 mm. screen. More than half of it is foreign material, granite and greenish quartzite predominating. Much of the basalt is weathered. Seventy-three per cent of the total may be called loess.

ward to 1,380 feet A.T. about 100 feet below the nearby scarp base. On the south side the gravel deposit extends up to about 1,420 feet A.T.

Two distinct ridges with a constructional sag between them stand on top of the bar where crossed by the highway. Cuts through these ridges show complicated depositional structures that are very difficult to explain in detail, though they clearly record a tumultuous flow of water out of Palouse Valley and back up Willow Creek across the top of the bar. The material is mostly angular basalt cobbles and pebbles with subordinate amounts of coarse, black or gray basalt sand. There are some definite strata and irregular lenses of fine sand and silt. Some of the finer material is cleanly sorted and stratified, but some is charged irregularly with grains of coarse basaltic sand.[1] Beds of well-stratified gray sand in one place dip eastward up Willow Creek at angles ranging from 22° to 35°. These beds rest with steep contacts on the edges of the nearly horizontal stratified basaltic gravel. Close to the contact, the stratified sand dips 35°. About 6 feet farther east, out in the sand, the dip is 30°; and in another 6 feet to the east, the dip is 22°. Overlying this foreset sand bed is a subangular basaltic gravel, also foreset and dipping about 20° to the east. Minor distortions in the sand foresets show that there was some slumping during aggradation, or soon afterward, but the inclination of the beds as a whole is not due to any slumping. Vigorous shifting currents in the great eddy back into the mouth of Willow Creek are shown in the diverse character and stratification of the débris, in the steep contact of these dipping beds on nearly horizontal beds, and in the other erosional surfaces shown in the section. It must be remembered that all this is on top of the bar deposit, at least 100 feet above the floor of Willow Creek valley.

In the upper part of this valley, east of the entrance of glacial waters across the divide from the north, silt terraces are recognizable for a mile or so. This is the same pebbly and sandy silt already described for other valleys and contains bits of quartzite, granite, and

[1] This pebbly silt appears to be identical with that in the valleys upstream from the scablands. The extraordinary mechanism that produced the intimate intermixture of coarse and fine material seems to have operated at times during the deposition of this bar.

other foreign rock.[1] None of the material was found higher than 1,550 feet A.T. The nearest scarp base whose altitude is known is about 3 miles up the scabland gradient at 1,540 feet A.T. Certainly the bar described is a constructional form which blocked Willow Creek valley; certainly its total vertical range is 200 feet; certainly its summit was swept by strong currents out of the main scabland; certainly Willow Creek contained backwater at this time; and certainly the bits of fresh granite and unweathered basalt in the terraces farther up Willow Creek did not come down that drainage way.

All the valleys thus far considered are tributaries of Palouse River, though only one (Lancaster Creek) enters that river east of the scabland. All the streams immediately south of Union Flat Creek, except Willow Creek, enter the canyon of the Snake; and only through it do they open into scabland. The huge glacial river entered the valley of Snake River across a wide tract of scabland whose eastern edge is close to Riparia. Examination of the astounding features recording glacial backwater in the Snake and its tributaries will begin at Riparia and will be extended up that drainage way as far as it has been studied.

### SNAKE RIVER VALLEY AND ITS TRIBUTARIES

There are enormous mounded gravel deposits in Snake River canyon near the junction of Palouse River where the glacial flood entered from the north. They may be traced eastward up the northern side of the Snake almost to Riparia. The deposit is thickest (260 feet) near the entrance of the Palouse and thinnest (150 feet) at the eastern end, but there is no definite progressive lowering of altitudes up the valley for the 5 miles of its extent. One minor gulch draining from the northern wall of Snake River valley was dammed by this deposit to a depth of at least 100 feet and has subsequently cut a narrow ravine through the gravel. A comparison of the ravine slopes

[1] Two samples, collected a mile apart, one at 1,500 feet, the other at 1,550 feet A.T., show very similar characters. One is 75 per cent loess; the other, 67 per cent. Three and six-tenths per cent of the material in one is composed of particles larger than 1 mm; 7.4 per cent, in the other. Both contain vein quartz and greenish quartzite, one yielded a fragment of mica schist. Basalt particles dominate in the coarser screenings; some much weathered, some very little altered. In both there is an increase in the percentage of angular quartz particles down at least to the 0.147 mm. sieve.

and the broad rolling surfaces of the deposit shows how little the original topography of the gravel deposit has been modified by run-off.

The débris for this rude terrace-like deposit came from the north in considerable quantity directly over the eastern part of the basalt wall against which the deposit rests. In several places it filled, or nearly filled, pre-existing minor gulches descending this wall, making remarkable features that look more like very long narrow gravel fans than anything else. But each fill is traceable up to scabland surfaces at about 1,000 feet A.T., and each contains non-basaltic débris like that already noted in the valleys to the north. There is nothing like these gravel-filled ravines in the Snake Valley above the entrance of the scabland river.

In structure, this Riparia "terrace" is largely composed of the long deltaic type of foreset beds, and these foresets dip prevailingly up the Snake. There are erosional contacts between many members, very irregularly distributed throughout the deposit, and with no suggestion whatever of weathering of the eroded surfaces. A number of striking lens-shaped bodies of silt and sand are present, apparently fillings of channel pools. Some of these lenses are without a trace of lamination, though 18–20 inches thick.

The composition of the gravel is mostly basalt, though the percentage of non-basalt is higher than for the deposits already described north of the Snake. The non-basaltic material is almost all very well rolled, whereas that in the valleys farther north is commonly angular or subangular. Quartzite material in the plateau gravels is prevailingly bluish and greenish, whereas quartzite in the gravel of Snake River valley is prevailingly white and yellow. Very suggestive of the source of these well-rolled fragments of light-colored quartzite is their abundance in the Snake River gravel now in transit.

### ALKALI FLAT CREEK VALLEY

This stream enters Snake River about a mile upstream from the eastern terminus of the "terrace" just described. For several miles back from the mouth of this tributary valley, it is narrow and deeply cut into the basalt. For about a mile back from the Snake this canyon has a coarse basaltic sand deposit in terrace remnants, 75–100

feet above the Snake at low water. The sand contains a few pebbles of angular basalt. The only foreign pebble found was a whitish quartzite. The deposit is very well bedded in foresets, the beds dipping 14° in the plane of the section across and up the canyon. No other high terrace sand or gravel was seen farther up this valley. But at Hay, which is 12 miles from the Snake, and for 5 miles farther up-valley, are silt deposits to a maximum altitude of 1,290 feet A.T.—600 feet higher than the coarse sand deposit. The depth of the silt is slight and sections are poor, but the characteristic abundant grains of basalt and a few particles and pebbles of foreign material are readily recognized.[1]

Eastward, above 1,300 feet, the valley for several miles has basaltic gravel in the streamway, and much loess has been washed down the slopes and along the valley. Yet this combination of basalt and loess in transit does not yield the tiny angular particles and coarse grains of country rock disseminated through a loessial alluvium. And of course it does not possess foreign particles. No trace of the pebbly, gritty silt was found above 1,300 feet A.T. This figure is approximately the upper altitude of the loess scarps at the junction of Snake Canyon and scabland. This same statement can be made for almost all the other valleys thus far described, though some do not have the basaltic stream gravel from which the tiny particles might conceivably originate. Where basalt fragments are mingled with loess in valley alluvium, they are larger than in the silts, are all

[1] Four samples of this silt were collected at intervals along a 5-mile stretch of the valley floor, distributed through an altitude range of 300 feet. The coarsest material, stopped by the 1 mm. sieve, ranges from 7 per cent to 18 per cent of the total; while the finest part of the samples, the reworked loess, ranges from 61 per cent to 83 per cent of each total. Basalt dominates the 1 mm. and 0.71 mm. screens; most of it rusty in color and distinctly weathered, none of it really fresh. None of the basalt is rounded. Foreign rocks and mineral fragments, all angular, include feldspar, quartz, porphyry, granite, and quartzites of whitish, greenish, and brown colors. The feldspar grains are clean cleavage fragments that show no evidence of weathering under a magnification of fifteen. There is an abundance of quartz grains on all sieves down to 0.147 mm. Since 90 per cent or more of all loess sieved passes 0.147 mm., these larger quartz grains are probably a contribution from the glacial waters, not a local derivative. The weathered condition of most of the basalt in these samples seems to indicate local derivation from slopes of weathered basalt. Associated fresh rock fragments show that the basalt has not weathered since deposition. In analyses of silts considerably nearer the Snake, a great deal of the basalt content of these silts apparently is unweathered and surely is much less weathered.

weathered, and are commonly arranged in strata. Nowhere do they have associated foreign particles.

CENTRAL FERRY

This place is at the crossing of the Inland Empire Highway, about 25 miles by river above the mouth of the Palouse. Two tributary valleys enter the Snake here—Deadman Creek from the south and a small unnamed creek from the north. The mouths of both these tributary valleys contain remarkable mounded gravel deposits, well exposed in pits opened for highway material.

The deposit in the mouth of Deadman Creek begins 1,000 feet back from the Snake and extends about $\frac{1}{4}$ mile farther upstream on the east side of the valley. It constitutes a half-mound resting against the higher basalt slope. The top of the half-mound makes a bench-like form 80 feet above the creek. There are many basalt outcrops through the thin soil immediately above the flattish top. The deposit is composed chiefly of basaltic rubble, nearly 100 per cent angular, very irregularly bedded but possessing distinct foreset strata in places, the dip of which is away from the Snake and up Deadman Creek, or partially up this valley and partially toward the basalt slope against which the deposit rests. Several angular granitoid boulders occur in the deposit. Two of them are of a rather striking granodiorite porphyry, one being 3 feet in diameter. Boulders and cobbles of this rock, or a porphyry so closely similar to it in megascopic and microscopic characters that no distinction can be made, have been found in several main scabland channels on the plateau. They are all subangular. None of this rock has been found in the rounded pebbles and cobbles now being transported by Snake River.

No other deposits of gravel well above the stream have been found farther up Deadman Creek. But the peculiar silt deposit is conspicuous in many road cuts for 9 miles up the valley, and an unweathered erratic cobble was found half a mile east of Gould, nearly 14 miles from the Snake, and 1,215 feet A.T. The silt is imperfectly stratified and poorly sorted. It is full of tiny pebbles and large sand grains composed chiefly of basalt but containing a few nonbasaltic particles, unweathered. The silt deposit nowhere shows the

development of a soil profile or a weathered upper zone. This is true throughout the entire region. It is to be contrasted with the common exposure of both A and B horizons in cuts in the loess above the level of the silt.

From the terraces along the lower part of Deadman Creek it appears that the silt originally completely covered the valley floor to a depth of 10 feet or more. But it also extends up on the slopes above the terraces. Low spurs of basalt, projecting out into the terraces, carry about the same thickness of silt as is found in the terrace sections. And along the Inland Empire Highway up a tributary to Deadman Creek a few miles south of Central Ferry, the same silt with the same characteristic unweathered coarse material scattered through it and the same kind of imperfect stratification is exposed in many places. It reaches an upper limit of at least 1,200 feet A.T.[1] At this level the silty deposit lies on unusually steep slopes, 200 feet or more above the nearby valley bottom, and 600 feet above the Snake at Central Ferry. In other words, the silt is a mantle which originally lay on all slopes up at least to 1,200 feet here and whose thickness on the steep upper slopes is not much less than in the valley bottom. Its deposition was so recent that weathering has not affected it and slope wash has not removed it. Certainly there has been no deepening of the valleys east of the scabland since this mantle was deposited.

The tributary entering Snake River from the north at Central Ferry carries a mounded gravel deposit in its valley mouth. It also

[1] In the sample from this altitude, most of the coarser material is non-basalt. Schist, micaceous quartzite, green quartzite, intergrown quartz and muscovite, clear quartz, two kinds of granitoid rock, and two kinds of gabbroid rock are represented. One of the gabbro-like pebbles is clearly smoothed and striated on two sides. Some of the quartz grains contain intergrown bits of biotite projecting from the exterior. Much of the basalt in the sample is so much decomposed that it crushes in the tweezers, but some of the chiplike forms are so slightly altered that the knife leaves glistening steel marks on their sharpest edges. There are almost no rounded forms in the coarser grades obtained by sieving this sample. Angularity of shapes is very pronounced. The proportion of quartz increases in each sieve sample down to 0.147 mm., where it is 50 per cent. About 90 per cent of the total sample is loess.

In contrast with this reworked loess is the undisturbed loess about 100 feet higher. Ninety-nine per cent of it passed every sieve. The 1 per cent stopped on the sieves was composed of calcareous aggregates of the loess.

lies about 1,000 feet back in the tributary valley and on the east side. Similarly, also, it is a half-mound in shape and has a benchlike flattish top. The top is 60 feet above the adjacent stream channel and 185 feet above the Snake. It has been undercut by the tributary and gashed by several gullies, but there is no difficulty in differentiating these steeper erosional slopes from the original mound slopes.

The stratification of this deposit is very irregular. There is definite foreset bedding in some of the cobbly débris, the beds dipping up the tributary, or toward the east wall, or somewhere between these two directions. Most of the material is basaltic rubble, very little worn to 100 per cent angular. Sorting in certain parts has made a few beds of well-stratified sand; but the bedding in the sand itself is irregular, and contact of the sand lenses with angular cobbly débris is also very irregular. Some of the sand is foreset with the dip up the tributary, like that of the coarser débris. Subangular boulders of granite are present, and there is one of the porphyry. There are numerous well-rounded pebbles of non-basalt like those so common in the present gravel of Snake River.

The deposit shows weather-staining only in the upper few feet. The granite and diorite boulders are fresh and unstained. So are the more worn basalt fragments. Yet the associated very angular basalt boulders are stained on all sides. It seems possible to explain these weathered surfaces in the midst of unweathered material only by assuming that they were weathered when contributed. Their marked angularity then can only mean that they are locally derived, that they traveled a very short distance. They cannot be explained as berg-carried like the angular erratics, for none are glacially worn or abraded, none occur stranded with the erratic boulders at high altitudes or on gentle slopes, and no basalt has been found in the definite berg "nests" or patches of berg-laid till that are yet to be described. They could not have fallen into the foreset strata; they must have been carried with the rest of the débris back into the valley mouth.

No other gravel has been found farther up the tributary, except close to the valley bottom. But there are many sections showing the pebble-charged silt up to about 1,200 feet A.T. The black basaltic sand in many places is distinctly segregated into irregular pockets in which there may be as much sand as silt. They possess finger-like

projections which penetrate the surrounding silt in various directions.[1]

There are no sections in unconsolidated material for about half a mile upstream from the 1,200-foot contour crossing. At 1,300 feet A.T. is a good section of fine-textured material resting on basalt, but it is simply clean loess with hardly a particle of basalt and none of any other kind of rock in it.[2] Essentially every road cut farther along the highway for 25 miles—all of it well above 1,300 feet A.T.—shows the same characters. The upper limit of the anomalous silt is close to the altitude of the scarp bases where the enormous glacial river entered the Snake near Riparia.

Snake River valley itself at Central Ferry contains a gravel deposit as extraordinary as those back in the two tributary mouths. At the highway crossing, the valley floor is a mile wide. The river here is against the south wall and down in a trench about 60 feet deep. The mile-wide floor is the surface of a gravel fill. In it are cuts for $2\frac{1}{2}$ miles along the Camas Prairie Railroad, which closely follows the north side of the river. Seen from the highway bridge, the gravel in the north bluff of the trench is stratified, the beds prevailingly horizontal and composed of well-worn stream pebbles and cobbles, apparently simply a fill of Snake River débris, into which the stream is now trenching. But in the next few paragraphs the reader will find that this deposit is so highly abnormal that no large valley deposit ever previously described can be cited as even similar to it.

In the first place, the surface of the deposit rises 75 feet northward from the bridge to the mouth of the unnamed tributary from

[1] A sample was taken at 1,175 feet. Eleven per cent of it stopped on the 1 mm. sieve; 62 per cent passed all sieves. Several bits of rock foreign to this tributary were found among the particles exceeding 0.5 mm. in diameter. Most of it is basalt; some fresh; some rusty red in color and capable of being scraped into a clayey powder. Quartz grains were found on all the coarser screens, some of them with intergrown mica and feldspar. The quartz content increases from 10 per cent on the 0.5 mm. screen to 50 per cent on the 0.177 mm. screen.

[2] Contrast this sample taken at 1,300 with that of the pebbly silt at 1,175. Of the total, 2.5 per cent was stopped by the first 9 screens and 10 per cent by the .074 screen; 87.5 per cent passed all screens. The material on the first 9 screens consisted of much decayed basalt with irregular shapes that can best be described as "crumbs." Though angular, they differ notably from the cleanly fractured outlines of most of the basalt in the pebbly silt. A very few quartz grains were found on the coarser screens. The color of the finest material in both samples is almost the same.

the north. Here the main valley deposit constitutes a broad ridge lying across the minor valley mouth, though 40 or 50 feet lower than the top of the semimound back in this tributary mouth. The tributary has cut a narrow gulch about 25 feet deep through the gravel ridge, just as Willow Creek, Kamiache Creek, and other streams have done where they enter scabland. It has also widely aggraded its own floor immediately upstream from the blocked entrance. The surface of the broad gravel fill, traced down the Snake Valley, descends gradually to river level in about 2 miles, so that there are no cuts along the railroad. The same descent of the surface and consequent vanishing of the cliff along the river occurs upstream also. This is no terrace profile, either across the valley or along it. The surface is decidedly undulatory, but not in terms of gullies or ravines. It is a huge, flattish half-mound, and it has been deposited as such, not eroded to this shape.

In the second place, the structure shown in the railroad cuts is almost without exception *foreset up the Snake*. The foreset structure is shared by the coarse gravel, the fine gravel, and the sand in the deposit. Not every stratum of each textural grade shows it, but most of the material (except lenses and strata of silt) does. Only two places were seen east of the bridge where the gravel has downstream foresets, and perhaps eight or ten lenses of the finer constituents possess downstream dip in their layers. The foreset gravel is arranged in nearly horizontal members, with as many as four such members in vertical sequence. One member may be only a foot in thickness; another may be 10–12 feet thick.

In the third place, the deposit, though dominantly of far-traveled Snake River gravel, contains almost unworn cobbles and boulders of local basalt. They are grayish, brownish, greenish, and reddish on the outside, typically like the weathering colors shown along joints in quarries of the region. They break readily along other joint cracks, showing similar colors when opened. Yet the material with which they are associated is essentially all fresh. The percentage of basalt among the cobbles and pebbles of the deposit is small, but it is classifiable into rounded fragments that are clean, bright, and hard, and angular fragments that are dull and decayed. The angular fragments average larger than any of the rounded basalt pieces. Some

boulders are 3, 4, and even 5 feet in maximum length, very much exceeding in size any fragments of non-basalt or any fragments of well-worn basalt.

The deposit contains coarse sand or fine gravel which is almost black in color because so highly basaltic, and is apparently identical with that found in the backwater silts described in this paper. It also contains strata and lenses of buff-colored silt or very fine sand, from a few inches to a few feet thick. Both occur from the top well down toward the base of the railroad cuts. But the intimate intermingling of the particles of coarse black sand in the loess-textured silt was not seen.

Basalt as a constituent of this Snake Valley deposit at Central Ferry is much more prominent in the highest portion, near the northern valley wall. In road-cuts half-way between Central Ferry and Purrington, it is almost all basalt sand and gravel, with minor foresets which dip both upstream and downstream. The deposit here is piled up against the basalt cliffs like the two half-mounds in the tributaries entering at Central Ferry. Yet it is only the more pronounced upper part of the great flattish half-mound crossed by the highway at Central Ferry.

In the fourth place, there is a definite relation between the altitudes on this deposit and those at which the two tributary valley mouth deposits stand. Each of these deposits is piled up approximately 50 feet higher than the top of the nearest part of the great deposit in Snake Valley. Yet the increase in altitude of the great deposit northward across the width of Snake Valley brings the base of the northern tributary's deposit higher than the top of the Deadman Creek deposit. Each of these minor half-mounds obviously was built with reference to the nearby surface of the larger, lower deposit.

It is too much to expect that the foregoing very brief description can convey an adequate picture of these deposits in their setting. But if one will rapidly review the more striking features, such as the strictly localized half-mound constructional forms, the altitudes involved, the slight amount of erosion and weathering of the deposits, the foresets dipping upstream in three diversely oriented valleys, the large fresh boulders of granite and porphyry, the very angular,

stained basalt boulders embedded in unstained gravel and sand, the associated silt mantle on higher slopes with coarse grains of fresh basalt sand and pebbles of non-basalt disseminated through it, the 650-foot vertical range of this silt, the close approximation of the upper limit of the silt with that of the scabland where the Snake enters that tremendously scoured topography, and the similarity of most of these features with those from Kamiache Creek south to the Snake, he must admit that the explanation of a great rush of berg-charged glacial water up the Snake could ask for little else in the way of record. Can any other explanation or group of explanations take care of all these items? Does any reader know of valley deposits elsewhere than on the Columbia Plateau which have similar features?[1]

VICINITY OF PENAWAWA

This town is 7 or 8 miles up the Snake from Central Ferry bridge. The station altitude is 613 feet A.T. Deposits similar to those just described are here also. The deep valley of Penawawa Creek up at least to 1,100 feet A.T. contains the silt, with bits of fresh basalt and small pebbles of granite and other rocks foreign to the valley.[2] A

[1] The glacial-flood hypothesis itself encounters difficulties in explaining this Central Ferry deposit. (1) The silt lenses tell of minor local pauses in current followed by renewed transportation of gravel up the valley. (2) A borelike wave moving up the valley would be succeeded by deepening water, and with increasing cross-sectional area of the reversed current; a decreasing velocity would maintain the same advancing wave. The gravel deposit, therefore, should be the product of only the vanguard of the flood. Could it conceivably have been so produced? (3) It seems questionable that a body of water adequate to make this up-valley current could, in terms even of the scabland-flood hypothesis, be poured across the Palouse-Snake divide near Riparia. (4) Most of the foresets are composed of Snake River débris, which by this explanation represents a pre-existing deposit torn up and relaid. Since no remnants of such older deposits show beneath the anomalous foreset-bedded gravel, and since it is impossible that older gravel could be completely torn out and its material rearranged and redeposited where the preflood gravel lay, it must be assumed that (*a*) a large basal part of the half-mound is of older débris, (*b*) the reworked material with added basaltic boulders covers the edges of the earlier deposit, partially filling the trench or inner valley of the preflood Snake, and (*c*) the dominant basaltic material constituting the apex of the broad mound is a contribution from valley-wall erosion and covers the surface of the older deposit.

[2] Six per cent of the sample from 1,100 feet A.T. was stopped by the 1 mm. sieve; 64 per cent passed all sieves. Though there is plenty of non-basalt in the three coarser grades, they are dominantly of very sharply angular basalt particles. Some particles are from two to five times as long as they are wide or thick. The sharp edges of many are hard enough to take a steel mark from the knife blade. The proportion of quartz, also with marked angular outlines, increases to 60 per cent in material stopped by the 0.177 sieve.

quartzite pebble was found at about 1,200 feet A.T., but nothing except loess and basalt were found above this altitude. The silt is thicker in the lower part of the valley, and in some sections the deposit is clean basalt sand. No mounded form of gravelly débris was seen in the valley mouth; but about a mile west of the railroad station, in a small gulch at Penawawa High School, there is a remarkable deposit of this character with foreset strata dipping back into the tributary out of Snake Valley. This deposit extends at least 325 feet above the Snake and originally must have nearly filled the gulch. Some parts of the deposit are composed wholly of cobbly débris, very little worn, all basalt, all with vari-colored surfaces of the original joint-plane weathering, all rudely foreset back into the little gulch. Other parts are composed of well-sorted, worn pebbles, also foreset toward the walls of Snake River canyon, i.e., back into the gulch. Thirty feet below the top of the deposit is a large cobble of the porphyry, definitely in place as a part of the extraordinary fill. In the lower part of the sections is 20 feet of coarse basaltic sand with shallow foresets arranged much like a huge pile of overlapping saucers tilted in the same direction. These dip out toward the Snake but surely are not a deposit of the local drainage. They record an eddy in some deep current flowing along the master valley. The débris of the fill is almost wholly basaltic and, though indubitably deposited by river currents, is quite unlike the normal Snake River gravel. The angular stained cobbles clearly were derived from the slopes and not from the bottom of the Snake Valley.

Cuts along a road up the south wall of Snake River canyon, 2 or 3 miles east of Penawawa, show the silt with numerous black basalt grains, some of them grouped in pockets, and with a few non-basalt pebbles and grains up to 1,200 feet A.T., about 600 feet above the river. No trace of such material was found above this altitude.[1]

[1] Two samples were taken here—one from the common phase of the pebbly silt at 1,200 feet and one from a pocket containing coarser material at 1,175 feet. The higher sample contained only 5.7 per cent of material larger than 0.5 mm.; while 9 per cent of it stopped on the 0.074 mm. sieve, and 73 per cent of it passed all sieves. The varied foreign material is unweathered and unstained, though the associated basalt particles are soft enough to scrape down readily under the knife blade. There is the increase in percentage of quartz with increase in fineness of material (40 per cent on 0.177 mm. sieve) common to the coarser screenings of all this silt. The sample from altitude 1,175 feet has

At the foot of this road-grade is Rice Bar, a large fill of gravelly waste in Snake River valley. It is half-mound in shape, its profile along the contact with the valley wall being a rude vertical arc, each end coming down gradually almost to river level. Its surface against the wall is nearly 100 feet higher than at the brink of the 20-foot river scarp at its northern margin. In this scarp the material is largely or wholly of well-rolled Snake River gravel. The section does not show structure. The surface of Rice Bar is undulatory: One undulation is a broad ridge paralleling the Snake and standing about 25 feet higher than the surface between it and the valley wall to the south.

In the highest part of the deposit, essentially against the basalt walls, a 10-foot excavation shows only well-stratified, black, fine gravel, more than 99 per cent basalt. It contains no slope-wash débris of loess or of fragments of weathered basalt, such as should occur if this half-mound were an alluvial fan in part. It is a river deposit, though certainly not composed of typical Snake River débris or dating back to a time when Snake River canyon was 100 feet less in depth. In the dominance of basaltic sand on top and the dominance of Snake River gravel in the basal part, Rice Bar duplicates Central Ferry Bar.[1]

Another half-mound deposit, known as Miller Bar, lies a few miles downstream from Rice Bar. It is very similar to Rice Bar and also stands about 100 feet higher along the southern wall of the canyon than at the brink of its low river-cut cliff. Here, as in Rice Bar, its higher portion cannot be interpreted as talus or alluvial-fan ma-

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7.7 per cent of its material exceeding 1 mm. in diameter and has percentages ranging from 2.9 to 8.3 for the next eight finer sieves. Twelve and a half per cent was stopped on the 0.074 mm. sieve and 43 per cent passed all sieves. Chiplike forms of the basalt particles are common, some with hard edges and black color. Grains of quartz increase in proportion to other constituents through the first six sieves, amounting to 40 per cent on screen 0.177 mm. Grains of unweathered feldspar were found, also grains of quartz with hornblende and biotite partially embedded.

[1] The shapes of fragments are extraordinarily angular. They include sharp-edged chips, and splinters in which the conchoidal fracture surfaces are unworn. A few bits of quartzite and flakes of mica were found. Quartz grains are more common in the finer grades of this sand, but their percentage is low. Half of the sample was stopped by the 1 mm. sieve; about a half of the remainder was caught on the 0.71 mm. sieve; and about half of what passed this sieve was stopped by the 0.5 mm. sieve. Nothing reached the 0.104 mm. sieve.

terial. Several small tributary gulches from the canyon wall mouth along its upper margin, but between them the deposit is as high, or is even higher, though not of talus débris or with talus forms. There are talus piles everywhere along here for comparison, and they are easily distinguished from the high edges of these bars.

### ALMOTA CREEK

No mounded deposit constricting the mouth of this tributary was identified, but for a mile or so back the pebbly silt shows in road-cuts, associated with pockets and lenses of clean sand. The silt contains foreign particles and angular bits of fresh basalt. Along the serpentine road up the canyon wall, north of Almota, are numerous foreign pebbles and cobbles. They are most abundant close to 1,250 feet A.T. One flattish cobble was delicately but definitely striated on both sides. A large cobble of the granodiorite porphyry lies at about 1,000 feet A.T.

In the canyon of Little Almota Creek, a small erratic boulder at 900 feet A.T. is the highest record of glacial water found well back from the Snake.

### WAWAWAI CREEK

A deposit composed largely of basaltic sand has been opened on the lower slopes back in the mouth of this valley. Its summit is about 230 feet above the Snake. Only the upper 10–15 feet are exposed in the section. It has no characteristic Snake River débris, though foreign particles are present. Though now constituting a spur out into the valley mouth, it is probably but a remnant of the fill which is recognizable in rude terraces farther back up Wawawai Canyon. These terraces are composed of the pebbly and gritty silt, with associated well-sorted sand layers up to about 900 feet A.T. Unsorted or but poorly sorted silt occurs up to about 1,200 feet A.T. It contains pebbles of granite, gneiss, schist, quartzite, etc. A fragment of the porphyry was found at 1,160 feet A.T., and a three-foot boulder of coarsely crystalline basic lava still higher.

### ALPOWA AND SILCOTT

At Alpowa there is a definite half-mound, though considerably cut away by the Snake along the upstream side. It has an alluvial

fan built out on it, but below the fan it is composed of worn river gravel. Its summit is 175 feet above the Snake.[1]

Alpowa is just within the deep canyon-like valley that leads out of the capacious Lewiston structural valley. Silcott is just above the head of this canyon, at the debouchure of Alpowa Creek. East of Silcott for a mile extends a mid-valley hill of gravel, its summit 830 feet A.T., about 130 feet above the Snake on the north and 90 feet or so above an empty channel or fosse on the south. The elongated hill has an undulating surface which descends gradually westward and southward, being highest next to the river. If there is anything definitive in its form, it is a bar, not a mid-valley terrace remnant. The only exposure is in an old pit at the east end. The material here is horizontally bedded, well-rounded unweathered river gravel, 24 per cent Columbia River basalt. In it are large angular weathered boulders of basalt.

The drainage of Alpowa Creek, entering the Snake at Silcott, comes entirely from the basalt plateau. Yet about 10 miles up its valley, 1,125 feet A.T. and 300 feet above the summit of the mid-valley hill, is a granite boulder 2 feet in diameter, embedded in pebbly silts along the Inland Empire Highway. These silts, a few feet in thickness, contain bits of granite, quartzite, etc., though their pebbly content is chiefly angular basalt fragments.[2] One striated non-basalt pebble was found. This deposit constitutes rude terraces in the creek valley below the 1,000-foot contour crossing. It establishes the

[1] Foreset beds dipping up Snake River valley have been seen at three other places than those described: (1) at Ridpath station, about midway between Riparia and Central Ferry, (2) 4 miles east of Almota, and (3) 2 miles east of Wawawai. They were noted only from a passing train. No foresets dipping downstream were seen.

[2] The collection was made a little below 1,300 feet A.T. in the highest road-cut which showed the pebbly silt.

Twenty-two per cent of the sample was stopped by the 1 mm. sieve. Fourteen per cent passed all but the 0.079 mm. sieve, and 42 per cent passed all sieves. The material retained on the 1 mm. sieve consisted of a few pebbles and a large number of coarse grains. The pebbles (maximum diameter, 1½ inches) were all foreign and all unweathered. The coarse grains were almost wholly basalt. The knife-edge, applied under a magnification of ten, left shining streaks of steel on the sharp edges of some of these pieces. Other pieces yielded readily under the knife. Angularity of almost all fragments is very marked, though some of the larger foreign pebbles are subrounded. The proportion of bits of quartz increases from a few pieces on the coarsest sieve to 60 per cent on the 0.177 mm. sieve. As previously pointed out, this quartz is too coarse to be a derivative of the loess and surely does not come from the basalt.

presence of a water body up to this level and of berg-ice floating in that water. There were no local valley glaciers in the region. The only alternative sources for berg-ice are the headwaters of the Clearwater or the Salmon or the Snake itself. According to Lindgren,[1] the valley glaciers of the Clearwater Mountains were very limited in extent and did not descend below 4,000 feet A.T. The termini of such valley glaciers were from 50 to 100 miles farther upstream than this 1,300-foot ponding could possibly reach. According to Lindgren's map, only granite, quartz-monzonite, diorite, and gneiss constitute the glaciated tracts of these mountains. He says that the "granite shows great constancy in its petrographic character" and that the quartz-monzonite so closely resembles the granite that "the two kinds of rocks cannot easily be distinguished in the field." The erratic blocks distributed up to 1,300 feet throughout the Snake and its tributaries represent a considerable variety, megascopically, of granite and granitoids, gabbros, porphyries, fine-grained lavas, quartzites, argillites, slates, schists, and gneisses. The evidence of the porphyry has already been cited. It seems impossible that the berg-ice of the Snake Valley ponding could have come from the glaciers in the Clearwater Mountains. And bergs from glaciers at the head of Snake River would have traveled 575 miles from the mountains of western Wyoming and eastern Idaho to reach the ponded tract.

### VICINITY OF THE SNAKE-CLEARWATER JUNCTION

Lewiston, Idaho, and Clarkston, Washington, are built at the junction of the Clearwater and Snake rivers in the lowest part of the Lewiston syncline. In the bottom of the structural valley the two rivers at their junction have probably not eroded more than 200 feet,[2] and this they have done concomitantly with the erosion of Snake Canyon westward from Alpowa and Silcott.

The town and orchard tracts of Clarkston, Washington, are located on a sloping terrace-like fill composed of sand and gravel with prominent deltaic foresets. The deposit is a broad semimound resting against the southwest wall of the valley, its surface descending

[1] W. Lindgren, "A Geological Reconnaissance across the Bitterroot Range and Clearwater Mountains in Montana and Idaho," U.S. Geol. Survey, *Prof. Paper 27* (1904), p. 116.

[2] I. C. Russell, "Geology and Water Resources of Nez Perce County, Idaho," *U.S. Geol. Survey, Water-Supply Paper 53* (1901), p. 70.

from the highest part roughly 50 feet per mile, down the valley, across the valley, and up the valley. Its upstream margin has been cliffed by undercutting of the Snake. This semimound is an impossible shape for a terrace top or a delta top, and the slopes are far too steep for either feature in a valley as broad as the Snake. There is no tributary drainage entering here from the higher land to the southwest, to which it might be referred as an alluvial-fan deposit.

Pits in several places afford some idea of the nature of the material. Two different deposits underlie the tract: an older stained and weathered gravel beneath an unweathered younger deposit in which Columbia basalt constitutes about 80 per cent. A group of exposures along the Snake in the southern part of the Clarkston semimound show dominantly dark unweathered fine angular gravel (Fig. 2) whose beds are inclined at various angles, some as steep as 35°. These foresets dip south, up the Snake. There are many erosional unconformities in the deposit here, developed by shifting bottom currents during deposition and promptly filled with beds which conform to the erosional surfaces, like a stack of shallow bowls in section, the upper bowls progressively broader and flatter. The fine gravel contains large lenses and strata of yellowish silt and fine sand. Some of the silt is well sorted; some is charged with irregularly dispersed angular bits of rock; chiefly fresh basalt. In most respects, this texture duplicates the pebbly silt already described from Kamiache Creek valley south to the Snake and back in the valleys of half a dozen tributaries of the Snake.[1] Three small angular boulders and one angular pebble of the porphyry were found here.

[1] Though this mixture of coarse sand, fine sand, and silt is apparently identical in structure with the high-level pebbly silts, it differs somewhat in texture and composition. The percentages stopped on the first five sieves range from 9 to 12.4, making the cumulative curve much steeper for these grades than in any pebbly silt thus far examined. Correspondingly, the quantity passing the 0.074 mm. sieve is unusually low—only 29 per cent of the total. Most of the material on the coarser sieves is very angular basalt. Most of this basalt is unweathered, a condition to be expected, for the sample was obtained nearly 30 feet below the surface. But a considerable number of basalt fragments are well weathered, crumbling between the tweezer blades or becoming clayey material when scraped with a knife. Such weathering must have occurred before the deposit was made. There are no large pebbles of foreign rock in the sample, though nonbasalt is present. The proportion of quartz on each of the sieves increases from 1 or 2 per cent on the 0.71 mm. sieve to 50 per cent on the 0.177 mm. sieve. The finest material is essentially loess, like all the high-level occurrences.

The summit material of the Clarkston semimound is composed of the familiar pebbly silt and coarse black sand. The silt is found up to 1,050 feet A.T. on the steep slope descending from the Clarkston Heights bench to the Clarkston semimound. A "nest" of angular bouldery fragments of non-basalt was found at 1,175 feet A.T. on

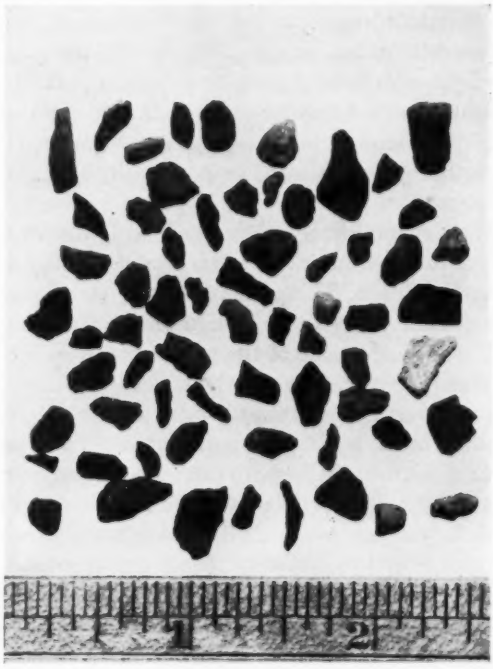

FIG. 2.—Chips, splinters, slivers, and broken-rounds from the Clarkston half-mound

the eroded slopes of the bench, the material quite unlike the high-level gravel, and apparently a berg deposit made subsequent to the erosion of the bench.

The residential section of Lewiston is built on a gravel terrace which rises steeply 50–100 feet just south of the business section. Numerous excellent exposures show dominant deltaic foresets dipping northward, down the Snake or across the Clearwater. The ma-

terial is grayish or yellowish but not markedly weathered. It is typical rolled river gravel and distinctly unlike the gravel, sand, and silt of the Clarkston deposit. Exposures in the top of the terrace, however, show none of this underlying gravel. There are many shallow cuts and every one examined shows tan-colored silt and fine sand, irregularly stratified, containing tiny angular fresh basalt pebbles and a few non-basalt fragments, but totally lacking in rounded pebbles and in gravelly strata. In the eastern part of the city, 20 feet or more of the light-colored sand and silt, with a few pockets and strata of coarse black sand, are shown resting on the older gravel beneath. A few layers of the sand and silt have numerous tiny angular pebbles and large angular grains of fresh basalt, partially arranged in stratiform fashion.

The surface of this younger deposit of the Lewiston terrace descends steeply to the south into a minor erosional valley which leads westward to the Snake. This steep margin is crenulated, and the width of the minor valley is variable in terms of the lobes and reentrants. The southern slope of the valley, however, is gentle, and its ground plan is regular. There are no crenulations in it. The flood hypothesis considers this southward slope, with lobes into an older erosional valley, as the lee face of a bar. The valley is in part a fosse.

A mile or so south of the Lewiston terrace, the surface rises in the southern wall of the structural valley to the Lewiston Orchards tract, 1,400 feet A.T. and 700 feet above the Snake and Clearwater. Cuts along the highway up this wall show the silt and sand with abundant fresh basalt particles and some angular erratics to 1,225 feet A.T.[1] A pocket of angular erratics was found at 1,200 feet,[2] and an angular granite fragment 16 inches in diameter at 1,225 feet. This is the highest known exposure of the pebbly silt in the Lewiston region. It

[1] Little foreign material was found on the 1 mm. screen, but it is varied. Both weathered and unweathered basalt occurs, the knife-sharp edges of some basalt splinters and chips being perfectly fresh. The quartz content of each sieve sample mounts from less than 5 per cent on the coarsest (1 mm.) to 60 per cent on the 0.147 mm. screen.

[2] Thirty-three per cent of this sample was stopped on the first two sieves, and 32 per cent is the sum of that passing all sieves or stopped by the finest. The larger pebbles ($1\frac{1}{4}$ inch maximum) are all of non-basaltic rock. Granite, schist, and quartzite are represented. The material on all the coarser sieves is largely of very angular fragments, in which chips, splinters, and slivers of rock are common. Quartz constitutes 40 per cent of the material on the 0.177 mm. sieve, most of it being in exceedingly angular fragments.

has basalt particles essentially as fresh as those taken from the pebbly silt embedded in the Clarkston halfmound. Yet these particles occur nearly 500 feet higher on the walls of the Lewiston Valley and only 4 or 5 feet below the surface, while the material from the pit was taken about 30 feet below the surface.[1]

### TAMMANY HOLLOW

Tammany Creek is a small perennial stream entering the east side of Snake River, nearly 5 miles south of the junction of Clearwater River. It has a valley flat a quarter of a mile or so in width for several miles above the mouth. But in a fashion strikingly reminiscent of Kamiache Creek and similar to many other tributary valleys already considered, the mouth is blocked by a barrier of gravel nearly 100 feet high. Through this the creek has cut a very narrow gorge at the contact of the gravel deposit and the northern basalt wall.

This barrier is a broad ridge of gravel nearly a mile long, lying almost wholly outside the mouth of Tammany Hollow, but extending

[1] We may admit that these erratic cobbles and boulders up to 1,300 feet A.T. in Snake River valley came from the Cordilleran ice sheet, yet deny that the episode was an extraordinary flooding. It is pertinent now to inquire whether their known distribution in the Snake drainage may be a test of the flood hypothesis. The alternative explanation, it will be recalled, is that scabland was made by glacial drainage of several different Pleistocene epochs, that the earliest of these occurred when the streams east of scabland "were flowing at levels far above their present courses," and that notable valley deepening by normal stream erosion has occurred during this succession of Pleistocene epochs. The stranded erratics of Snake River drainage are highest above the valley floors nearest its entrance into scabland. It must be assumed under the alternative explanation that this height, about 800 feet, is the sum of two quantities: (1) the depth of the earliest flooding and (2) the amount of subsequent valley-deepening. By this alternative the two quantities might be estimated at 100 feet and 700 feet, respectively. How far up this early Pleistocene Snake River valley, 700 feet less in depth than now, would a ponding extend which was only 100 feet deep at the mouth of the Palouse? Only as many miles as the gradient in feet per mile is contained in 100. The present gradient of the Snake here is about 2½ feet per mile. It was probably steeper at an earlier stage in valley development. A 100-foot ponding today would produce backwater for only 40 miles up the Snake. Is it possible that the gradient of the preglacial Snake was so low that the erratics stranded at 1,225 feet A.T. above Lewiston, 70 miles upstream from the scablands, were transported here during the earliest rational flooding?

To wait for another glacial epoch will be equally difficult, for we are committed to a considerable deepening of Snake Valley in the meantime, larger fractions of the total deepening being logically assignable to the earlier interglacial epochs. Indeed, it is im-

completely across that valley mouth from one headland to the other. The summit of the barrier is nearly 1,050 feet A.T., about 250 feet above the Snake. From the summit the slope toward the Snake is relatively gentle, that facing up Tammany Hollow is steep. There are no gravel terraces in the Hollow. At mid-length this barrier consists chiefly of long deltaic foresets. They dip in that utterly anomalous fashion noted repeatedly in this study, directly out of the Snake Valley and up the tributary valley. The material is dominantly fine, dark-colored gravel; the pebbles, exceptionally angular and chiefly of Columbia basalt (Fig. 3). But there are plenty of cobbles and boulders in these foresets. Most of the boulders are composed of much-decayed basalt buried in the almost perfectly fresh finer débris.

To complete the picture, one needs only the unsorted pebbly silts and some stranded angular erratics farther back up the gulch behind this barrier. There are no good cuts to show the presence of the silt. Only loess and alluvium and basalt ledges show. But an angular granite boulder, a foot and a half in diameter, and an angular cobble of the porphyry were found about a mile east of the barrier dam.

In summary, the relatively broad Lewiston synclinal valley, at the head of the canyon that extends from Alpowa almost to the Columbia River, contains the following aberrant depositional features:

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possible to get bergs very far up the Snake at any time by this mechanism. It is impossible to carry the upper limit any farther upstream than the intersection of that high-water line (about 1,300 feet A.T.) with the vanished floor of the early Pleistocene Snake River canyon. It *cannot* intersect the present floor. By this hypothesis it is equally impossible for any but the latest high-water line to intersect the present canyon floor, though as the gradient of the deepening Snake River valley decreases, each succeeding flood will reach somewhat farther up the valley. The upstream limit of foreign berg débris must, by this hypothesis, *be a descending gradient.* This it is not. Only a Snake Valley system of full present depth at the time of glacial backwater will provide the existing horizontal upper limit of berg-carried débris to the distances up-valley that are already known.

Final establishment of this thesis will be made when the Snake and Clearwater valleys have been studied where the upper-limit contour crosses them. In most of the valleys entering scabland from the east, this flood-carried débris has already been found up to, but not above, the crossing of the proper contour.

Other serious objections to this alternative hypothesis include (1) the lack of widening of the Snake Valley during this deepening, the lack even of notable slope wash or gullying, which would remove the earliest high-line records; (2) the lack of weathering in the high-lying stranded erratics; and (3) the lack of decay in the tiny chips of basalt so abundant in the silt at all altitudes below the highest limit.

A. Silts:

1. Pebbly and coarse sandy silts occur in the bottoms and mainly on the slopes of both main and tributary valleys, extending up nearly or quite to 1,300 feet A.T., 600 feet above the Snake at the Clearwater junction.

2. The abundant particles of coarse sand and small pebbles occurring in the silts are irregularly disseminated or in pockets; rarely stratiform.

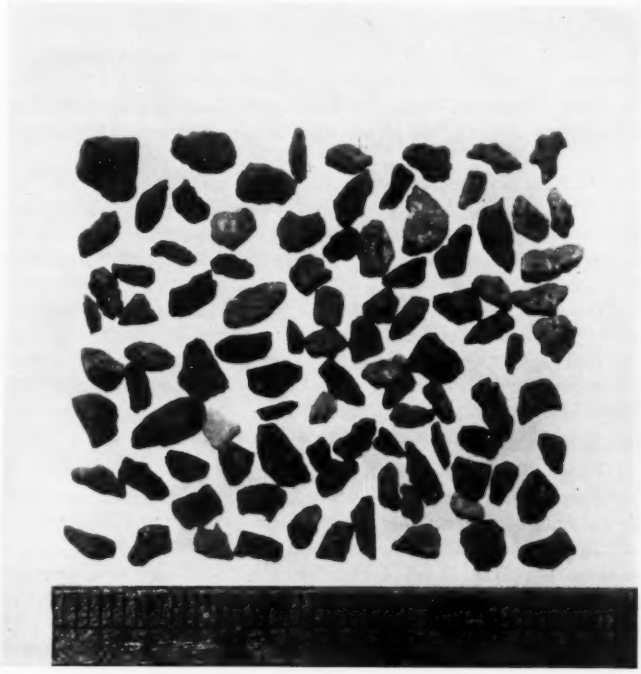

FIG. 3.—Chips, splinters, slivers, and broken-rounds from Tammany Hollow Bar

3. These larger particles are dominantly of basalt, but there is everywhere a small percentage of non-basaltic material back in valleys draining wholly from basalt country rock.

4. Most of these particles are very angular and are unweathered. The lack of weathering is as conspicuous at the upper limits as in the valley-bottom occurrences.

5. The material of this silt is identical with the loess of the plateau top and the gentle valley slopes of the region.

B. Sporadic boulders:

1. Angular boulders of unweathered granite and other non-basalts occur stranded alone, or embedded in silts, or in gravel, at all altitudes below the upper limit of about 1,300 feet A.T. The granodiorite porphyry boulders and cobbles in this class probably are derivative from the scabland glacial rivers.

2. Angular boulders of local basalt, commonly much weathered, occur in gravel deposits of fresh material in the valley bottoms.

C. Gravel deposits:

1. Mounds and half-mounds, 100–200 feet thick, composed largely of basaltic gravel, occur on the floor of Snake Valley.

2. One such mound, elongated with the main valley, once constituted a complete barrier at least 100 feet high across the mouth of a good-sized tributary valley.

3. Structure of these mounds, where exposed, contains foresets, inclining in various directions with the varying slopes of the mound. There is one marked case of foreset dip up the Snake, the location being nearer the middle than the wall of the valley. In another case, the dip is directly out of the Snake and up a tributary.

4. The material of these mounds is largely unweathered; but contains fragments up to boulders in size which are much weathered.

5. There are seams and lenses of light-brown silt in the gravel, some with the same irregularly disseminated sharp bits of basalt and non-basalt that the silts themselves carry.

D. Relation to the regional topography:

1. The silt, foreign boulders, and gravel deposits are very young.

*a*) They lie on erosional slopes where any notable amount of valley deepening or widening would inevitably remove them.

*b*) The foreign boulders, and the unweathered pebbles in the silt, lying high up on these slopes, cannot date back to early stages of valley development.

*c*) Gravel deposits, which once were complete barriers to tributary streams, are still almost intact.

2. The silts, foreign boulders, and gravel deposits record some abnormal episode of flooding.

*a*) Unweathered materials ranging from fine basalt sand to large foreign boulders cannot be distributed through this amazing vertical range, the fine débris essentially as a blanket on all gentler slopes, by any ordinary sequences in valley history.

*b*) The absence of such records everywhere above 1,300 feet A.T. indicates this altitude as the approximate upper limit of the flooding. Above it the surface deposits are loess or basalt rubble. The basal portions of the loess commonly have angular fragments of weathered basalt, but no foreign material and no fresh basalt particles. Nor is the loess rudely stratified; nor does its rare basaltic débris occur in pockets.

3. That abnormal episode of flooding involved strong currents in all valleys.

*a*) Extreme angularity of the basaltic débris records pounding and battering, rather than ordinary abrasive grinding, to produce it.

*b*) Erratic material in long, narrow canyons could hardly have floated back unless the episode were very long continued. Current transportation seems necessary if the time interval be short. Other evidence, already cited, indicates brevity of time involved.

4. The current of that abnormal flooding was, in every valley, exactly the reverse of the normal drainage which made these valleys and which now reoccupies them.

*a*) Erratic boulders were berg-floated back into tributaries from the Snake and were berg-floated up the Snake from the scablands.

*b*) Erratic débris was mingled in suspension with silt (the rehandled loess) and deposited in many cases without assorting.

*c*) Foresets of the gravel deposited during this episode in significant places dip back up the valleys. Eddies will not account for such dip.

*d*) If a mechanism be provided by which the flood water could have come from the headwaters of the various valleys, or of even only the major ones, there certainly should be an ascending upper limit with increasing distance upstream. If it is 1,300 feet at Riparia, it should be considerably above 1,300 feet at Lewiston and Clarkston, 65 miles farther upstream. No such rise is indicated. There is only loess and weathered basalt above this level.

5. There was *in no case any outlet* for these currents in the direction in which they flowed. The only conceivable cause for these reverse currents, in the light of all data now in hand, was an extraordinary volume in the scabland rivers and an extraordinarily rapid rise to the maximum of that volume.

[*To be concluded*]

# FROST HEAVING[1]

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## ABSTRACT

This paper describes a laboratory investigation of problems connected with frost heaving. The principles developed have applications in geology, plant physiology, and engineering.

Pressure effects accompanying the freezing of soils are due to the growth of ice crystals and not change in volume. Pressure is developed in the direction of crystal growth, which is determined chiefly by the direction of cooling. Heaving is often greater than can be explained by expansion. It is due to the segregation of water as it freezes, more water being drawn up by molecular cohesion. The chief factors controlling segregation and excessive heaving are: size of soil particle, amount of water available, size and percentage of voids, and rate of cooling. The pressure developed is limited by the tensile strength of the water.

More water may be concentrated in the surface soil through ice segregation followed by thawing than in any other way. Segregation causes shrinkage cracks below if the supply of water is limited or the soil very impermeable.

Differential heaving is due chiefly to differences in soil texture and in the amount of available water, but differences in the kind and amount of soil cover are also factors.

## INTRODUCTION

Heaving and subsidence of the surface, due to freezing and thawing of soils, have caused much damage in cold climates to hard-surfaced roads, foundations, and also to certain agricultural crops. The uplift of the surface soil has usually been attributed to the change in volume that takes place when water freezes, the expansion being upward because of less resistance in that direction.

Several observers have called attention to the columnar ice crystals that uplift pebbles and clods of earth as evidence of a force which could not be attributed to expansion in volume, one of the earliest and clearest statements having been made by Volger.[2] This evidence, however, has failed to convince most writers, and, so far as I can ascertain, almost no experimental work has been done on the problems connected with frost heaving.

Experiments carried out by me on cold nights during the winter

[1] Presented in part at the New York meeting of the Geological Society of America, December 28, 1928.

[2] G. H. Otto Volger, "Über die Volumveränderungen, welche durch die Krystallisation hervorgerufen werden," *Annalen der Physik und Chemie, Vierte Reihe* (herausgegeben zu Berlin von J. C. Poggendorff, 1854), Band XCIII, pp. 232–37.

of 1914–15 indicated that the pressure effects accompanying freezing are due to the growth of ice crystals and that excessive heaving is to be explained by the segregation of water as it freezes. This segregation was obtained in clay but not in sand. After publishing brief descriptions of some of these experiments together with certain conclusions,[1] I laid the investigation aside temporarily, because of lack of facilities for obtaining and maintaining low temperatures.

In March, 1927, suitable low-temperature apparatus was placed at my disposal by Mr. E. W. Allen, district manager of the Frigidaire Corporation, and I began an investigation to determine the factors involved in excessive and in differential frost heaving. I am indebted to my colleagues in the science departments of the University of South Carolina for suggestions and assistance during the progress of the investigation. Since November, 1927, the work has been carried on through co-operation with the United States Bureau of Public Roads.

The experiments have been conducted in a room having nearly constant temperature. The freezing cabinet, cooled by expansion coils that extend across the top, contains a box of sand, suitably insulated, in which the soil containers are packed so that freezing takes place from the top down just as in the ground. A thermograph records the temperature of the air in the cabinet, and a thermocouple is used to determine soil temperatures below the surface. In most tests the soils are frozen in cylindrical pasteboard containers, made impervious with paraffin or shellac. At the end of a test, the container is quickly cut away from the frozen soil cylinder, which may then be examined, measured, and photographed.

In some tests gallon containers have been used, but for most purposes the standard quart cartons, 17 cm. high and 8.5 cm. in diameter, have proved more satisfactory, as a larger number of simultaneous tests can be run.

Since the ice in some soils is formed in part of water drawn from the water-table, this condition was reproduced in experiments. Holes were punched in the bottoms of cartons and covered with filter paper before introducing the soils; then the cartons were set

[1] "The Growth of Crystals under External Pressure," *Amer. Jour. Sci.*, Vol. XLI (1916), pp. 544–45; "Ice Forming in Clay Soils Will Lift Surface Weights," *Eng. News-Record*, Vol. LXXX (1918), pp. 262–63; "Surface Heaving Caused by Segregation of Water Forming Ice Crystals," *ibid.*, Vol. LXXXI (1918), pp. 683–84.

in sand kept saturated with water. Collars fitting snugly around the lower part of the cartons prevented the wet sand from coming in contact with the dry sand in which the tests were packed.

## EXCESSIVE FROST HEAVING

The maximum amount of frost heaving that occurs during cold winters in the northern part of the United States is not known, very few accurate observations being available. An uplift of 6 inches has been observed in New York State, but the maximum is probably two or three times that amount.

When water present in the voids of a soil freezes without the introduction of additional material, the amount of heaving is necessarily limited by the change in volume that accompanies freezing; and, since the water-content of the average soil is seldom as much as 50 per cent while the depth of freezing in the United States seldom exceeds 2 or 3 feet, the amount of heaving under these conditions should not be more than 1 or 2 inches, for the expansion in volume of water on freezing is only about 10 per cent. Moreover, in most soils having a high water-content a considerable percentage of the water does not freeze.

Surface heaving appreciably in excess of that which would result from the expansion in volume of the water contained in the soil means that additional material must have been introduced as a result of the freezing. Laboratory experiments under favorable conditions give a surface uplift of over 60 per cent of the depth of freezing. Excessive heaving, in the laboratory, is always accompanied by the introduction of additional water and the segregation of some water to form masses or layers of more or less pure ice. Such ice, while usually solid and transparent, has a satiny luster and fibrous structure because of the orientation of the tiny ice prisms normal to the cooling surface (see Figs. 1, 6, 8, 12, and 18). Similar segregated ice has been found in the ground when excavations have been made in investigating frost heaving.[1]

## FACTORS INFLUENCING FROST HEAVING

In the investigation described in this paper an attempt has been made to determine the effect of different factors on the freezing of

[1] Lewis B. Wyckoff, "Some Observations on Effect of Frost in Raising Weights," *Eng. News-Record*, Vol. LXXX (1918), p. 628.

water in soils, especially of the factors involved in segregation and excessive heaving, by varying one factor at a time while the others are kept as nearly constant as possible.

### SIZE AND PERCENTAGE OF VOIDS IN SOIL

The size of the capillary spaces in soils is an important factor, for it determines the height to which water may be lifted above the water-table by surface tension. Other things being equal, the height to which water rises varies inversely as the diameter of the capillaries; in extremely fine material it may be 3 m. (10 ft.) or more, but most very fine soils have a high percentage of colloids, the effect

FIG. 1.—Frozen clay cylinder with sand layer in middle. *A*, frozen clay; *B*, sand; *C*, ice; *D*, unfrozen clay.

of which is to decrease permeability and prevent water from rising as high as it otherwise would.

A relatively thin layer of coarse sand will stop the upward movement of water by capillarity and thus prevent the formation of ice layers. This was illustrated by an experiment. A carton with perforated bottom was filled with tightly packed dry clay except for a layer of standard Ottawa sand, 2.5 cm. thick, in the center. It was placed in sand kept saturated with water and compacted several times a day until, at the end of four days, it became slightly moist on top; then it was subjected to freezing from the top down for four days. The result is shown in Figure 1. Because of lack of water very little segregated ice formed above the sand, and in the sand there was insufficient ice to bind the grains together. The cylinder was photographed on its side with only part of the carton removed in

order to avoid complete slumping of the sand. Below the sand and resting on unfrozen clay a layer of practically pure ice, 2.5 cm. thick, was formed.

In previous papers attention has been called to the fact that water does not freeze as readily in very small capillary spaces. In Figure 2 the ice lenses extend for about 1 cm. obliquely downward into unfrozen clay. Since cooling took place from the top down the isothermal surfaces should be nearly horizontal.

FIG. 2.—Spiral ice lenses in frozen clay cylinder. Unfrozen clay below.

In material having very small pore spaces a portion of the contained water remains unfrozen after long exposure to a temperature several degrees below freezing. The accurate determination of the percentage of water freezing in certain types of soil is a difficult problem. Previous investigators[1] have used small samples of soil (usually about 25 gm.); freezing has been rapid, heat being abstracted in all directions; and, usually, the sample has been agitated. These conditions are not conducive to ice segregation; therefore it was necessary to modify the methods hitherto used.

The soils were frozen in a modified dilatometer consisting of a glass cylinder with tightly fitting top having a long-drawn-out stem (see Fig. 3). Meta-xylene was placed on top of the soil, and the rise of this liquid in the glass stem gave a measure of the change in volume. Xylene was substituted for the ligroin, previously used, because the loss through vaporization was less during long experiments and it did not congeal at the low temperature required. The

[1] G. J. Bouyoucos, "Classification and Measurement of Different Forms of Water in the Soil by Means of the Dilatometer Method," *Mich. Agric. College, Tech. Bull. 36* (1917), p. 48; A. M. Wintermyer, "Percentage of Water Freezable in Soils," *Public Roads*, Vol. V, No. 12 (1925), pp. 5–8.

apparatus has a capacity of about 300 cc. Because of the size of the ground-glass joint between the cylinder and its top, much trouble was caused by leakage; but, after considerable experimenting, it was found that a perfect seal could be obtained with a little paste made of powdered Castile soap and water.

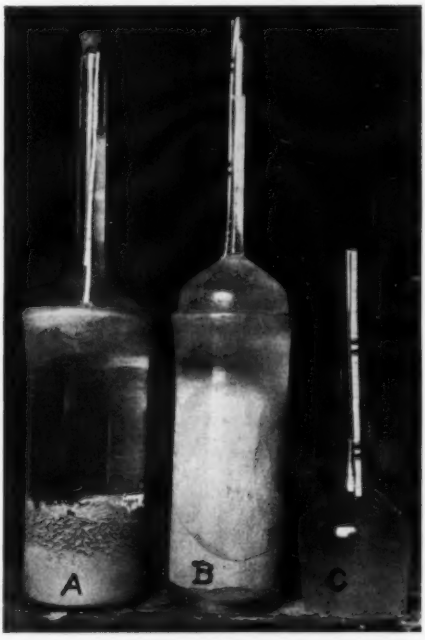



FIG. 3.—Dilatometers containing frozen clay and frozen lump of clay

The apparatus was buried in sand with only the top exposed and the air temperature reduced to between $-12°$ and $-15°$ C. Freezing was continued until the volume became constant. The temperature reached at the bottom of the cylinders was between $-5°$ and $-7°$ C., depending on the length of the tests, which varied from two to five days. The apparatus was tested by freezing 100 cc. of boiled distilled water, and during every test a blank was run in order to correct for the decrease in volume of the xylene.

When clean sand, or other material having relatively large voids, is saturated with water the change in volume on freezing indicates that all of the water is converted into ice; but in every test with clay, quartz flour, and other material having a large number of extremely small voids the increase in volume was less than that which would have occurred if all the water froze.

In obtaining accurate results with clay and similar material all air must be carefully eliminated. It is impossible to pack a uniform mixture of clay and water in a cylinder without including some air. When holes were bored in the bottom of a cylinder so that water could enter dry clay from below, the free air was driven upward out of the clay, but a few hours later adsorbed air separated out, forming cracks and bubbles.

Powdered clay was stirred into boiling water to form a thin mixture which was poured into a cylinder. Excess water was removed by siphoning and evaporation, the material being jarred and tamped occasionally to prevent development of shrinkage cracks. On freezing, segregated ice layers were formed in the upper part of the clay, but in the lower part shrinkage cracks ranging in width up to 3 mm. developed. Since air might have been absorbed during evaporation, a similar test was run in which the excess water was removed by evaporation under a vacuum with sulphuric acid as absorbent. Freezing resulted in no cracks, but the lower part of the clay shrank away from the glass (see *A* in Fig. 3). This indicates that a vacuum may be formed by the shrinkage of clay due to removal of water to form ice masses during freezing. The per cent of water present was determined by drying the clay after the close of the experiment.

A cylinder weighing 255 gm., cut from a lump of somewhat indurated Cretaceous white clay containing 20.4 per cent water as it came from the mine, was placed in xylene in a dilatometer and cooled as before. A layer of fibrous ice, 0.5–1.0 cm. in thickness, formed on top of the clay, but very little ice was visible elsewhere; shrinkage cracks developed in the lower part (see *B* in Fig. 3).

Another clay cylinder from which half the water had been removed by evaporation gave similar results except that the ice layer was thinner. It was necessary to delay freezing until xylene had displaced the air which partly filled the voids in the clay.

In all of these tests the change in volume indicated that for each 100 gm. of dry clay about 6 gm. of water failed to freeze. The amount of water present seems to make little or no difference in the results, but definite conclusions should not be drawn without more data.

In tests made on large fragments of porous porcelain saturated with water about seven parts of water failed to freeze for each one hundred parts of the solid present. The failure of part of the water to freeze is determined by pore space rather than size of particle, for the pieces of porous porcelain behave in other freezing tests like large particles. The water that does not freeze is probably adsorbed water, and adsorption is a surface phenomenon.

### SIZE OF SOIL PARTICLE

Size of the soil particle is one of the most important factors controlling segregation of water during freezing, and the effects are clearly distinguishable from those due to size and amount of pore space. Early in the investigation it was observed that excessive frost heaving occurred on clays but not on clean sands. Even when the sand is so fine that water will rise in it to a height of 20 cm. or more, segregation and excessive heaving do not occur.

In an attempt to ascertain the maximum size of particle that would permit segregation, clean quartz was ground in a ball mill and sized by screening. The quartz dust passing a sieve with 200 wires to the linear inch (maximum particle size about 0.07 mm. in diameter) when tested under the most favorable conditions gave only the faintest evidence of segregation (see Fig. 4). After regrinding in an agate mortar and bolting through silk, segregation was slightly increased. Microscopic examination showed that the average diameter of the quartz particles was probably between 6 and 10 microns.

Because of the difficulty of grinding quartz fine enough and sizing it accurately, chemically precipitated crystals of substances with behavior similar to quartz were substituted in some of the tests. Precipitated barium sulphate,[1] very uniform in size, having an

[1] The barium sulphate used in this investigation and also the lithopone (71 per cent $BaSO_4$ and 29 per cent ZnS) and kadox (ZnO) were obtained from the research laboratories of the Du Pont de Nemours Co. through the courtesy of Mr. Marion L. Hanahan.

average particle diameter of about 2 microns and a maximum diameter of 3 microns, gave well-defined segregation under favorable conditions of cooling (see Fig. 5), but no segregation under unfavorable conditions. In materials having a particle size of about

FIG. 4

FIG. 5

FIG. 4.—Quartz flour passing 200-mesh sieve, showing slight segregation of ice near top; lower part of cylinder unfrozen.

FIG. 5.—Barium sulphate with ice layer at lower limit of freezing; faint bands of segregated ice above. Iron weight on top.

1 micron or less segregation took place without difficulty. The results obtained with lithopone, which has a particle size of approximately $\frac{1}{2}$ micron, and with kadox, having a probable particle size of $\frac{1}{4}$ micron or less, are shown in *A* and *B* of Figure 6.

All of the tests were carried out under exactly the same conditions. The dry material was tightly packed in cartons, having per-

forated bottoms covered with filter paper, which were then stood in sand kept saturated with water. As the water rises by capillarity through very fine material, driving the air upward, the small particles are rearranged and more closely compacted so that the wet material below becomes separated from the dry material above by an air space large enough to stop the ascent of the water. Therefore, it is necessary to tamp the material occasionally with a heavy iron cylinder, and it takes several days for the water to reach the top.

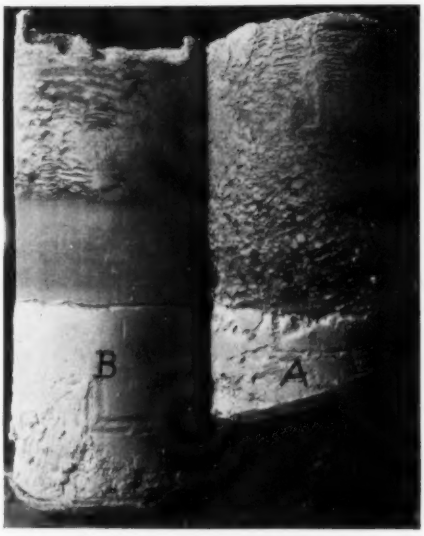



FIG. 6.—Frozen cylinders of (*A*) lithopone, with particle size of $\frac{1}{2}$ micron; and (*B*) kadox, with particle size of $\frac{1}{4}$ micron. Note layer of pure ice 5cm. in thickness.

During the process of freezing the water ascends much more rapidly or it would not be possible to build up such thick layers of ice.

The sedimentary clay used in these tests consists largely of particles having apparent diameters of between 1 and $\frac{1}{2}$ micron, but it is very difficult to distinguish the ultimate particles from aggregates. Part of the clay is much finer. A few flakes of mica ranging up to 0.1 mm. are present, and more rarely quartz grains having diameters up to 0.01 mm.

Tests were run on different mixtures of pure clay and standard

Ottawa sand. This sand is very uniform, the grains having an average diameter of about 0.7 mm., and when well packed has 32.9 per cent voids. Mixtures containing slightly less than 20 per cent clay by weight would have just enough clay to fill the voids between the sand grains so that the size of the pore spaces would be the same as in pure clay. On freezing no segregated ice could be seen in mixtures containing less than 30 per cent clay, and even with 50 per cent clay the amount of segregation is small as compared with tests made on pure clay. The results obtained by freezing sand-clay mixtures

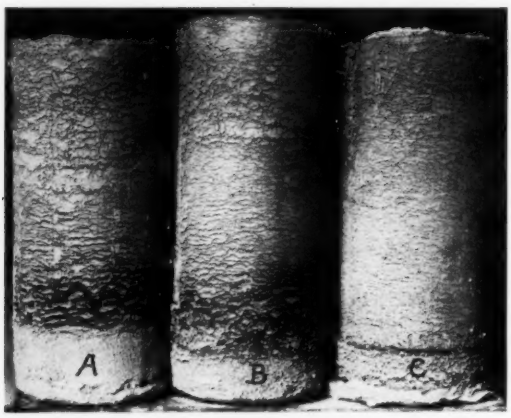



FIG. 7.—Frozen cylinders containing (*A*) 50 per cent, (*B*) 40 per cent, and (*C*) 30 per cent clay mixed with sand.

containing 50, 40, and 30 per cent clay, respectively, are shown in Figure 7.

The tests were prepared by weighing and thoroughly mixing the dry materials which were then tightly packed in cartons with perforated bottoms. The cartons were stood in sand kept saturated with water until the mixtures became wet on top; then they were removed and the material further compacted by tamping and evaporation. When shrinkage was nearly complete the cartons were again placed in the wet sand, and, after the mixtures were wet on top, they were frozen from the top down.

The influence of large particles in preventing segregation during freezing is probably to be explained as follows:

During the growth of a thick ice layer in soil the freezing isotherm does not move downward; as water, drawn up through the capillaries, crosses this isothermal surface it is converted into ice, and the overlying material is uplifted through a distance equal to the thickness of the ice. In order to maintain the position of the freezing isotherm relative to the unfrozen soil, the heat conducted away through the frozen material must balance the heat conducted up through the unfrozen material together with the heat brought up by the ascending water and the heat set free in converting water into ice.

The freezing isotherm cannot advance as rapidly in water as in the minerals found in soils, for it is a poorer conductor of heat, has a higher specific heat, and also much heat has to be removed in converting water into ice. Hence, if ice crystals in growing downward come in contact with the top of a large soil particle and begin to surround it so that the temperature at the top of the particle drops below freezing, then the temperature of the bottom of that particle will reach the freezing point before the water with which it is in contact. Therefore, freezing takes place in part outward from the surfaces of large mineral particles that are in contact with water, and not merely through the downward growth of ice crystals as in pure water.

When a growing ice crystal closely approaches a soil particle, the water separating them is gradually reduced to a very thin film and further growth of the crystal in this direction can take place only as molecules of water enter this film. If the soil particle is relatively small so that the molecules do not have far to travel through the film, and if growth is relatively slow, so that they have time to enter between the ice and the particle, then the growing crystal will exclude the particle; but if the particle is relatively large and if freezing is relatively rapid, the particle is gradually surrounded by the ice. To build up a layer of ice, which consists of many prismatic crystals, the capillaries supplying the water must be closely spaced.

The space occupied by a horizontal glass rod buried in clay was elongated in a vertical direction to form a large open space by the

growth of an ice layer. Even the hole left by the removal of a rod was similarly elongated (see Fig. 8).

Another experiment affords confirmatory evidence in support of the theory outlined above. A porous battery cell was placed bottom side up in a shallow vessel of water and surrounded with insulating material so that it could be cooled from the top down. When the temperature of the cell was above freezing and the air in contact with the exposed upper surface of the cell only a little below freezing, a layer of fibrous ice was formed on top of the cell; but when the air temperature was low, so that the freezing isotherm passed rapidly down into the cell, very little ice formed on its surfaces, and the cell did not break.

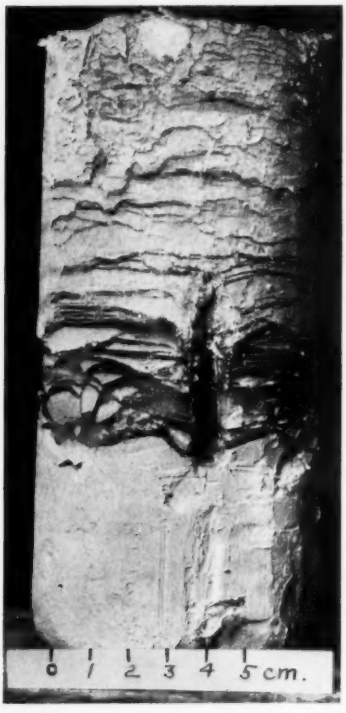



FIG. 8.—Frozen clay cylinder, showing elongation of hole by growth of ice layer.

The battery cell is similar to clay in having closely spaced capillaries through which water may be fed up to growing ice crystals and in which only part of the water freezes; but it is similar to coarse material in preventing the formation of ice layers within. The failure to break can hardly be attributed to the greater tensile strength of the cell, for, as will be shown later, the pressure developed by growing ice crystals may be very great.

Porous brick, cement, and stone even when saturated with water usually offer considerable resistance to fracture at low temperatures, especially if heat is conducted away in only one direction. This is probably because they are so thoroughly fused or cemented together that they behave as a single solid and not as an aggregate of small

discrete particles. If flaws, cracks, or large cavities are present they may, under favorable conditions, be enlarged. Repeated freezing and thawing, especially when cooling takes place from all directions, tends gradually to cause disintegration.

### WATER-CONTENT OF SOIL

The amount of heaving that takes place when soils freeze is limited by the supply of available water; not merely the water present in soils before freezing, but also the water that can be drawn from points below the depth of freezing. The effect of moisture content on segregation was determined through a series of tests on samples prepared with different percentages of water.

TABLE I

| Test No. | Water at Bottom of Clay Cylinders | |
|---|---|---|
| | Per cent before Freezing | Per cent after Freezing |
| 1 | 29.0 | 20.8 |
| 2 | 25.8 | 19.9 |
| 3 | 24.2 | 18.3 |
| 4 | 23.2 | 18.3 |
| 5 | 19.9 | 16.8 |
| 6 | 16.8 | 12.3 |
| 7 | 9.5 | 7.7 |

Dry powdered clay was packed in cartons with perforated bottoms and weighed. The cartons were then stood in wet sand until the clay was saturated, after which the excess water was removed by slow evaporation together with compression and tamping. This method prevented the development of shrinkage cracks. The cartons were weighed daily until their weights indicated that the desired moisture content had been reached, when they were sealed, top and bottom, with paraffin to prevent further loss. The clay cylinders were about 15 cm. high and 8.5 cm. in diameter. The tests were buried in sand and frozen from the top down, after which samples were taken from the bottom of each cylinder and the moisture content determined. The results are shown in Table I.

The small irregularities shown in the table may be attributed to slight differences in the rate of freezing and in the height of the

cylinders as well as the difficulty of obtaining uniform samples. Samples from near the tops of the cylinders were useless for comparison because of the presence of segregated ice layers in tests 1, 2, 3, and 4. These ice layers had a maximum thickness of 3 mm. in test 1, and were somewhat smaller in the other three. Tests 5, 6, and 7 showed no visible evidence of the segregation revealed by the moisture determinations. Proof that the segregation was due entirely to freezing, and not to previous treatment, was obtained by freezing check samples with the ends of the clay cylinders reversed.

A cylinder of clay, cut from a lump at the mine and imbedded in paraffin without loss of moisture, was tested in the same way. Samples taken after freezing showed a moisture content at top and bottom of 34.76 and 19.06 per cent, respectively.

When freezing tests were made on closed systems from which air and additional water were excluded, the clay reabsorbed the water from the ice layers soon after melting, but in tests where air and water were not excluded water often stood on top of the clay after thawing.

Another experiment sheds some light on the movement of ground-water during freezing. A cylinder 7 cm. in diameter and 15 cm. high, cut from a lump of somewhat indurated clay, was placed in a carton with perforated bottom and surrounded with paraffin. The bottom of the carton was kept in saturated sand while the clay was being frozen from the top down. As obtained from the mine the clay contained 20.46 per cent water. When the cylinder had frozen to a depth of 11 cm., with heaving of 4 cm. (see *C* in Fig. 11), a sample of frozen clay was obtained 8 cm. from the top between two ice layers, and another was taken in the unfrozen clay about 7 cm. above the bottom. The moisture content of these samples was 20.68 and 20.64 per cent, respectively.

The thoroughly consolidated clay used in these experiments has extremely small capillaries and the colloidal content is relatively low; therefore, water would rise higher above the water-table than in most soils. While the upward movement of water during freezing is very slow, there must be considerable frictional resistance to its passage, so that, other conditions being equal, frost heaving decreases as the depth of the water-table increases. However, the

effect of a low water-table could be in part compensated by slower cooling.

### RATE OF COOLING

The formation of segregated ice layers during the freezing of soils is favored by a very slow lowering of the freezing isotherm; therefore, the most favorable places for the development of ice layers are close to the surface and near the lower limit of frost penetration. Needle ice, or frost columns, develop at the surface of moist, clayey soils when the temperature of the ground immediately below the surface remains above freezing, while the air temperature is below freezing. They do not form when previously chilled soils are rapidly frozen during a sudden drop in temperature.

In material barely fine enough to permit segregation, such as the barium sulphate previously described, well-defined layers of ice were obtained only near the top and at the bottom of the frozen material. The absence of well-defined ice layers near the top of the barium sulphate shown in Figure 5 was due to rapid conduction of heat by a 4-lb. iron weight that rested on it during the experiment. The development of ice layers was entirely prevented by packing the carton of barium sulphate, together with the wet sand in which it stood, in ice for eighteen hours to bring the system to 0° C., and then transferring it quickly to the sand box, which had been brought to approximately the same temperature. The freezing isotherm moved rapidly downward as the air temperature was about −15° C. Pure clay, when treated in this way, was not appreciably affected, and ice layers developed as usual.

The rapidity with which soil temperature changes depends on many factors, the most important being the temperature gradient. Dark-colored soils absorb and radiate heat more rapidly than light-colored soils. Other things being equal, the rate of cooling is determined by the thermal conductivities and specific heats of the different materials present, the former being most important. Soils containing much organic matter change temperature more slowly than soils made up chiefly of minerals. The soil minerals differ from one another little in their thermal properties, but they differ greatly from air and water, the other constituents of soils. Therefore, at

temperatures above freezing, the rate of cooling of mineral soils is determined chiefly by the porosity and the water content.

When the soil temperature drops below freezing, another factor helps to determine the rate of cooling, i.e., the amount of water converted into ice. In sand this is limited to the interstitial water, but in clay much of the ice may be formed of water drawn from below during the freezing process. The difference in rate of cooling caused by difference in soil texture is well illustrated by the graphs

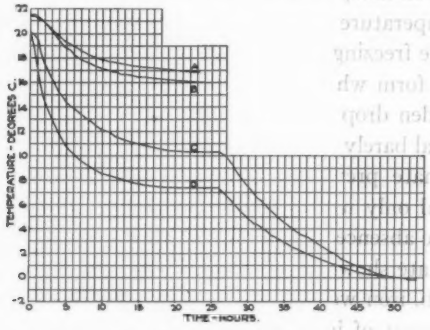



FIG. 9.—Cooling curves for barium sulphate and clay. *A*, barium sulphate; voids filled with air. *B*, clay; voids filled with air. *C*, barium sulphate; voids filled with water. *D*, clay; voids filled with water. Surface temperature 0° C. for first 26 hours, and then −21.3° C. until close of the experiment.

shown in Figure 9, which were plotted from data obtained as follows:

Barium sulphate and clay were packed dry in separate containers, which were then buried in insulating material in a constant-temperature room. The temperature at the surface of the material was maintained at 0° C., while at half-hour intervals readings were made to the nearest 0.05° C. on partial immersion thermometers inserted 8 cm. below the surface. The clay, containing 58.7 per cent voids, cooled down more rapidly than the barium sulphate with 69 per cent voids and reached a minimum temperature 1° lower.

The voids were next filled with water, which made possible somewhat closer packing, especially of the clay, as it is in large part composed of tabular-shaped particles, and then the experiment was repeated with the thermometers 7 cm. below the surface. Both

cooled down more rapidly than when the voids were filled with air; but the clay, containing 50.5 per cent water by volume, cooled down much more rapidly than the barium sulphate with 67 per cent water and reached a minimum temperature 3° lower.

After the temperature had become constant the surface temperature was lowered to −21.3° C. Under these conditions the clay cooled down more slowly than the barium sulphate, so that the graphs converge instead of diverging as before. This change in the relative rates of cooling was due to the formation of segregated ice near the surface of the clay, while little or no segregation took place in the barium sulphate. In this experiment the cartons were sealed at the bottom with paraffin. If water had been permitted to enter from below, the difference in cooling would have been even more noticeable.

### DEPTH OF FREEZING

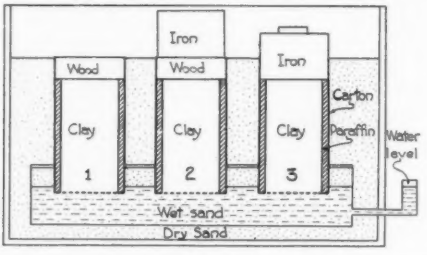
The amount of surface heaving is not proportional to the depth of freezing, but the latter is a limiting factor. The uplift may vary from zero in certain soils having a low water-content up to 100 per cent of the depth of freezing in instances where a layer of ice is formed at the surface with no freezing at points within the soil. The depth to which freezing extends depends on the minimum temperature reached, the length of the cold spell, the amount of heat present in soil and rock near the surface, the thermal properties of the soil and soil cover, and the changes produced in the composition and thermal properties of the soil because of segregation during freezing.

In laboratory experiments conducted at the same room temperature, the thick layers of segregated ice that mark the lower limit of freezing in clay developed at greater depth when the air in the refrigerator was kept at lower temperature. In similar tests in which the temperature of air in the refrigerator was kept the same, the depth of freezing was greatest when the room temperature was lowest.

The effect of soil cover as a factor in frost heaving was investigated experimentally. Three tests could be conducted simultaneously, and care was taken to secure comparable results. Cylinders 14.25 cm. high and 7.8 cm. in diameter were cut from moist clay of uniform texture and placed on filter paper in cartons with

perforated bottoms. The top and bottom of the cylinders were made smooth and parallel. Molten paraffin was poured between the cylinders and the sides of the cartons after a little clay had been tamped around the base of the cylinders to prevent intrusion of paraffin under them, as this would interfere with the entrance of water.

The cartons were placed in sand kept saturated with water during the experiment. A disk of wood, soaked in paraffin and weighing




| | Cylinder No. | | |
|---|---|---|---|
| | 1 | 2 | 3 |
| Load (gms. per sq. cm.) | 2 | 60 | 60 |
| Depth of freezing (cm.) | 8.5 | 10.0 | 10.8 |
| Uplift (cm.) | 3.2 | 4.1 | 4.5 |
| Ratio of uplift to depth of freezing | 0.38 | 0.40 | 0.41 |
| Time in freezing-box (hours) | 74 | 74 | 74 |
| Initial temperature of air (C.) | +16° | +16° | +16° |
| Final temperature of air (C) | −14° | −14° | −14° |

Fig. 10.—Apparatus for freezing clay cylinders when weighted with different materials.

0.1 kg., rested on one of the clay cylinders; a similar disk of wood and an iron weight totaling 3.1 kg. rested on the second cylinder; while the third was loaded with 3.1 kg. of iron (see Fig. 10). The depth of freezing and the ratio of uplift to depth of freezing were both greatest where the iron, a good conductor of heat, was in contact with the clay, and were least where wood was the only covering.

Similar tests were run on three clay cylinders, one of which supported 7.4 kg. of iron, while another with the same load had a wooden disk between the clay and iron, and the third had nothing on it. The depth of freezing and ratio of uplift to depth of freezing

were greatest where the iron rested on clay and least where the wood insulated the clay from the cold air (see Fig. 11).

The result of a test conducted in the same way, except that tightly packed clay was substituted for the cylinders cut from undisturbed clay, is shown in Figure 12. The clay supported a maximum load of 14 kg. (7 lb. per sq.in.), but it was necessary to remove

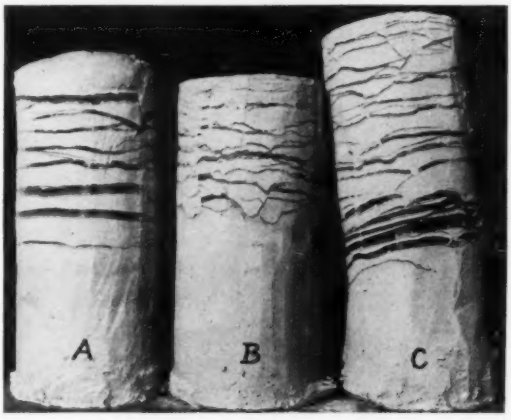



FIG. 11.—Clay cylinders frozen (*A*) under no surface load; (*B*) under iron weight insulated from clay by wooden disc; and (*C*) under iron weight in contact with clay.

some of the iron weights, because of lack of space, as heaving progressed. The uplift was 6.4 cm. and the depth of freezing 15 cm., the test being discontinued after 114 hours.

Tests show that at temperatures above freezing concrete pavements are better conductors of heat than clay, even when the latter is saturated with water.

### DIRECTION OF COOLING

It is commonly stated that frost heaving is upward because of less resistance to expansion in that direction, but this is not the true explanation. The pressure effects that accompany freezing are not due to change in volume, but to the growth of ice crystals. Crystals perform work in any particular direction by growing in that direction and overcoming the resistance to growth. The upward heaving that

accompanies the freezing of soils is due to the growth of ice crystals in a vertical direction, and this is usually determined by the direction in which heat is conducted away most rapidly, and by the availability of the water necessary for growth. The importance of the direction of heat conduction is illustrated in the experiments described below.

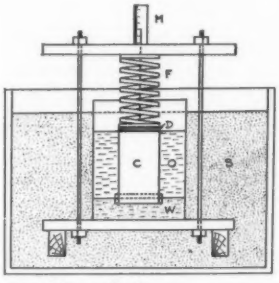
FIG. 12.—Clay cylinder frozen under 30 pounds of iron; shows excessive segregation of ice.

Mixtures of white clay and water in different proportions were frozen in thin glass test tubes, half of them being buried in sand so that freezing was from the top down, while the others were exposed so that freezing took place from the sides inward. All of the latter were broken, longitudinal cracks extending the full length of the test tubes. But where frozen from the top down none was broken, for the ice crystals grew only in a vertical direction. Alternating layers of clear ice and frozen clay were formed in the upper parts of the mixtures containing sufficient water, while shrinkage cracks developed below as water was withdrawn to build the ice layers above.

When test tubes filled with clear water are exposed to low temperatures they do not usually break unless there is a constriction in the tube, for the density of water decreases on cooling from 4° to 0° C., causing the coldest water to rise and come in contact with the downward-growing ice crystals. The clay not only prevents convection currents, but it also supplies capillaries through which water reaches the growing ice crystals.

Apparatus was devised in which it was possible to freeze clay cylinders from the top down under heavy vertical pressure and with

practically no lateral support (see Fig. 13). A cylinder 6 cm. in diameter, cut from hardened clay, was placed in a pasteboard tray with perforated bottom, and moist clay packed around it to hold it in place, a strip of adhesive tape, 2 cm. wide, having been wrapped around the bottom of the cylinder to prevent softening by the rapid absorption of water. The tray was placed in a large carton containing a layer of sand saturated with water, and oil was then poured in until level with the top of the clay. Carton and tray had been made impervious to water and oil by coating with shellac. The oil used was very viscous when cold, thus eliminating convection currents, yet it did not solidify at the lowest temperature reached.[1] A metal disk resting on the clay cylinder supported a stiff steel spring which was compressed against a board held in place by steel bolts. The amount of heaving and the pressure developed were indicated on a




C—Clay cylinder
O—Viscous oil
W—Sand saturated with water
D—Steel disk
F—Steel spring
M—Graduated scale
S—Dry sand

FIG. 13

FIG. 14

FIG. 13.—Apparatus for freezing clay cylinders under heavy vertical pressure and no lateral support.

FIG. 14.—Clay cylinder frozen under heavy vertical pressure with no lateral support.

[1] The "Humble 'D' Pale Oil" used in these experiments was furnished through the courtesy of Mr. J. W. Saybolt of the Standard Oil Company of New Jersey.

scale by a wire that extended from the top of the clay up through the spring.

When ready for freezing, the apparatus was buried in dry sand, the bolts having been tightened to give an initial pressure of 101 lb. Five hours after cooling was started, the pressure had dropped to 94 lb. because of slow failure of the cylinder, but fifteen hours later heaving had restored the pressure to 101 lb. The pressure increased gradually and reached a maximum of 140 lb. (2.3 tons per sq. ft.) seventy-six hours after the experiment was started. The pressure then remained constant as freezing had reached the bottom of the cylinder.

Examination of the clay cylinder (see Fig. 14) showed that heaving was largely due to the formation of horizontal veins of fibrous or columnar ice, ranging up to 0.5 cm. in thickness. The large vertical cracks, due to failure under pressure, contained little ice, being occupied chiefly by oil, but a few cracks, less than 2 mm. in width, were filled with ice.

In similar tests, conducted under less pressure, the clay cylinders showed no evidence of failure, and their appearance after freezing was essentially the same as the cylinders shown in Figure 11.

In another experiment clay was packed around heavy copper bars standing in a carton with perforated bottom. The carton was placed on wet sand and then the entire apparatus was buried in the sand box for freezing as in previous experiments. When removed a layer of fibrous ice, about a centimeter thick, was found surrounding the copper, the crystals being normal to the copper bars. Ice veins, ranging up to 2 mm. in thickness, branched off from the vertical ice layer and sloped steeply upward. Near the bottom of the carton a large mass of ice was formed which ruptured the carton and pushed part of the clay out into the inclosing sand.

The rapid conduction of heat by the copper caused the growth of ice crystals radially outward from its surface, and the pressure of these growing crystals ruptured the carton. In other experiments the cartons were never ruptured by the direct pressure of growing ice crystals because heat conduction, and therefore crystal growth, were limited to a vertical direction.

### EXTERNAL PRESSURE

The determination of the maximum force that may be exerted by growing ice crystals is an interesting but difficult problem. Appa-

ratus used in measuring pressures developed during frost heaving is shown in Figure 15. A carton containing a clay cylinder set in paraffin stands in sand kept saturated with water. The spring, supported by the clay, is compressed as heaving takes place, the amount of compression being measured as previously described. In most of the work a much heavier spring than the one shown in the photograph was used, and it was necessary to substitute steel plates

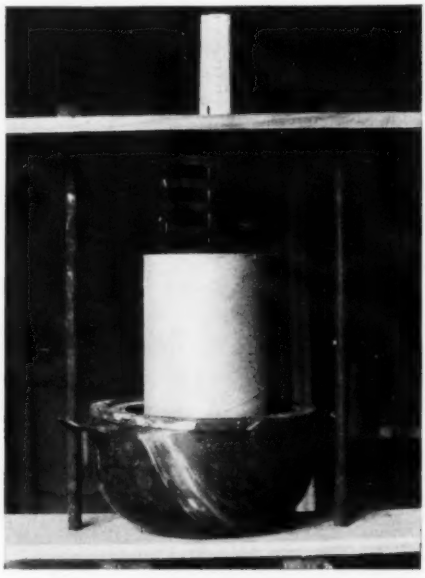


FIG. 15.—Apparatus for freezing clay cylinders under pressure

for the boards. While the clay was being frozen the apparatus was buried in dry sand up to the top of the carton.

In tests with this apparatus an indicated pressure of over 155 lb. per square inch (11 kg. per sq.cm.) was obtained before the side-wall of the carton was broken and pushed out near the bottom by softened clay. But the force exerted by the ice crystals must have been much greater, for in forming each layer of ice it was necessary to pry apart the clay cylinder and push it up against the friction of the paraffin in which it was imbedded. The clay, containing 20 per

cent water as it came from the mine, had a crushing strength of 200 lb. per square inch and a tensile strength of about 75 lb. per square inch, as determined in standard testing machines. Therefore, the maximum pressure developed by growing ice crystals in these experiments must have been well over 14 tons per square foot (14 kg. per sq.cm.).

In several tests with clay cylinders, vertical or steeply inclined planes of weakness developed against which the ice veins terminated abruptly, the growth of the veins causing local contemporaneous faulting along the flaws (see Fig. 16). While these faults were more common in tests made under high pressure, they also occur occasionally in tests made under no surface load. One of the clay cylinders showed an open vertical fissure when it was placed in the refrigerator. On freezing under pressure the fissure was filled with ice and contemporaneous faulting occurred along it because of the growth of horizontal veins. In some tests the lower parts of the clay cylinders were more or less shattered by the high pressures together with insufficient lateral support, and typical breccias were formed (see Fig. 17).

From these tests it is clear that excessive heaving may occur under surface loads of 150 or 200 lb. per square inch. The ratio of the uplift to the depth of freezing was much less, however, than that obtained under lower pressures. What would be the effect of much higher pressures?

If the frozen clay cylinders had no lateral support the pressures developed would be limited by the crushing strength of the material. Little is known concerning the crushing strength of ice, but apparently it increases somewhat as the temperature is lowered, and is greater in a direction parallel to the principal axis of the crystals than in other directions. Barnes and Mackay found the crushing strength of St. Lawrence River ice, at or near the freezing point, to be about 363 lb. per square inch.[1] Tarr and von Engeln, working with small blocks of pond ice at temperatures between 18° and 20° F., obtained a crushing strength of 1,000 lb. per square inch in the direction of the principal axis of the ice crystals and 350 lb. per square inch in a direction normal to them.[2]

[1] H. T. Barnes, *Ice Engineering* (Montreal, 1928), p. 214.

[2] R. S. Tarr and O. D. von Engeln, "Experimental Studies of Ice with Reference to Glacier Structure and Motion," *Zeitschrift für Gletscherkunde*, Band IX, Heft 2 (1915), p. 91.

Tests on frozen clay cylinders, made by the writer at temperatures near freezing, show that the crushing strength increases with the ice content, the maximum obtained being over 700 lb. per square inch. In all tests compression was applied in a direction parallel to

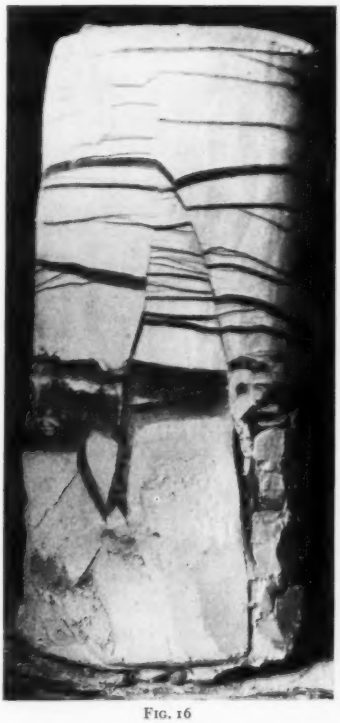


FIG. 16

FIG. 17

FIG. 16.—Clay cylinder frozen under heavy pressure, showing ice veins and contemporaneous faults.

FIG. 17.—Clay cylinder frozen under heavy pressure, showing breccia

the principal axis of the ice crystals, although in much of the material the ice crystals were too small to be visible to the naked eye.

During the freezing of soils the growing ice crystals are, as a rule, supported laterally by other crystals, so that the crushing

strength of the frozen material does not limit the load which may be lifted.

If the crushing strength of the unfrozen material is small, flowage will prevent the development of high pressures unless the material be confined. When water freezes in a confined space the resulting pressure may still be attributed to growing ice crystals, for the crystals develop hydraulic pressure by displacing the water with which they are in contact.

The effect of pressure on ice is slightly to lower the melting point, thus making necessary a lower temperature if crystal growth is to continue. Bridgman found that a pressure of 1,000 kg. per square centimeter lowered the freezing point to −8.8° C., while at 2,115 kg., where the freezing point is −22.0°, ice I and ice III are in equilibrium with each other as well as with water.[1]

Extremely high pressure would compact the clay, reduce porosity, and make it less permeable to water, but the entrance of water would be retarded rather than prevented, and, if the water entered slowly, the removal of heat would likewise have to be slow in order that ice segregation take place. High pressure may bring part of the clay particles into such close contact with one another that they conduct heat essentially as though the clay were partly composed of larger particles.

With the apparatus available it was impracticable to measure accurately pressures exceeding 150–200 lb. per square inch, but wet clays were frozen under much higher pressures by sealing them in wrought-iron cylinders, having a bursting strength of over 11,000 lb. per square inch. At first ice crystals grew radially inward from the exposed sides of the cylinder, but as the pressure increased and the wooden end plugs were slowly displaced, the ice crystals grew in the direction of least resistance, forming ice layers transverse to the axis of the cylinder. In the small iron containers cooling could not be limited to a single direction as in the other tests.

### DIFFERENTIAL HEAVING

Simple heaving of the surface, if uniform, causes little cracking of pavements or disturbance of structures; but differential heaving is very destructive. Local differences in soil texture and in the amount

[1] P. W. Bridgman, *Proc. Amer. Acad.*, Vol. XLVII (1912), pp. 441–558.

of water available are the two major causes of differential heaving. Differences in the kind and amount of soil cover, such as snow, paving and foundation material, vegetation, etc., are probably of importance. Local differences in insolation, air drainage in hilly regions, and differences in pressure due to static load are minor factors.

The effect of difference in soil texture is easily illustrated experimentally. A carton with perforated bottom was packed with fine sand on one side and clay on the other, by inserting a partition down the middle and gradually withdrawing it as the materials were rammed into place. On placing the carton in saturated sand, water rose slowly to the top displacing the air. Penetration was more rapid in the fine permeable sand, and therefore some water spread laterally into the clay. The carton was buried to the top in dry sand, so that freezing took place from the top down, but the bottom of the carton was kept in saturated sand throughout the experiment. At first only the clay was appreciably heaved, but later the sand also began to rise because unfrozen clay was being forced under it, and at the close of the experiment the bottom of the carton had been pushed out and to one side by the movement of the clay (see Fig. 18).

Another test was made in the same way except that, after saturating the sand and clay with water, the carton was sealed at the bottom by standing it in molten paraffin. On freezing, ice layers, 3–4 mm. thick, were formed in the clay, the required water being drawn from the clay in the lower part of the carton and also from the sand. At the close of the experiment, when the carton had been cut away, the lower half of the sand was so nearly dry that it flowed readily, and careful handling was necessary to prevent collapse until the photograph (see Fig. 19) could be taken. The upper part of the sand was thoroughly cemented by ice which filled all voids. Because of differential heaving, the soil cylinder was gradually tilted away from the clay side causing the wall of the carton to be pulled apart by tension.

Even though the soil texture is uniform, differential heaving will occur under favorable conditions where there are differences in the moisture content and especially differences in the position of the water-table relative to the surface. In experiments, the ratio of uplift to depth of freezing showed little decrease as the height of the soil column above the artificial water-table was increased.

EFFECTS OF THAWING

It has been shown that freezing may result in the concentration of excessive amounts of water in the form of ice at, or near, the surface. Melting of this ice usually takes place from the surface downward, so that removal by downward percolation is largely

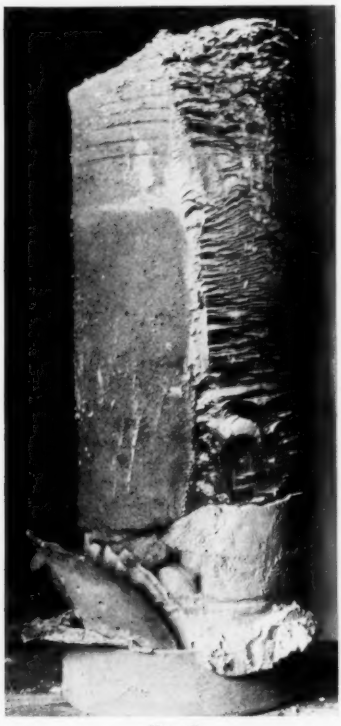


FIG. 18

FIG. 19

FIG. 18.—Frozen cylinder, half sand and half clay. Much segregated ice in clay but not in sand.

FIG. 19.—Differential displacement of cylinder, due to segregation of ice in clay but not in sand. Cavity caused by dislodgment of dry sand.

prevented until after most of the ice has melted. In this way much water may be concentrated under pavements where it could not accumulate as a result of the rapid melting of snow. These conditions are conducive to unequal settling, especially on slopes, and the resulting damage to pavements is, in my opinion, greater than that which

may be attributed directly to frost heaving. As the cracks may not develop except under heavy traffic and perhaps long after the freezing occurred, this cause of failure has not been appreciated.

### EFFECTS OF THAWING AND REFREEZING

Experiments have shown that freezing tends to concentrate water near the surface, and that segregation is accentuated by a high water-content. Therefore, prompt refreezing after thawing should result in greater segregation than on the first freezing. This conclusion was confirmed by thawing and refreezing clays which contained too little water to form ice layers on the first freezing.

A sudden drop in temperature after a spring thaw, which has left an excessive amount of water in the surface soil, is a condition favorable for segregation and excessive heaving. Bituminous macadam pavements seem to be especially susceptible at such times, as are also certain agricultural crops, such as wheat.

### SHRINKAGE OF CLAYS CAUSED BY ICE SEGREGATION

In every experiment in which no additional water entered the system, the withdrawal of water from the lower part of the container to build up ice layers above caused shrinkage in volume, and, usually, the development of tensional cracks. This shrinkage in volume furnishes an indication of the force with which water is drawn toward the growing ice crystals. Terzaghi has shown that the forces involved in the shrinkage of clay on evaporation are equal to the external pressure required to produce the same change in volume, and that these forces may be very great.[1]

In soils that are very impermeable because of high colloid content, the formation of segregated ice sets up tension in the material immediately below, and, since the stress is uniformly distributed, vertical cracks tend to develop and form a polygonal pattern (see Fig. 20). The cracks are gradually filled with clear ice, and, as freezing progresses, the ice-filled cracks advance downward to form a columnar structure, and this combined with the usual horizontal ice layers results in the cellular structure shown in Figure 21. The clay forming the interior of the cells does not freeze easily, and therefore

[1] C. Terzaghi, "Principles of Soil Mechanics," *Eng. News-Record*, Vol. XCV (1925), pp. 742–46 and 796–800.

the divisional walls of clear ice are gradually thickened at the expense of the clay in the interior. With decrease in the per cent of colloids the polygonal cracks become more widely spaced.

In tests requiring soils with a high percentage of colloids, bentonite and mixtures of bentonite with South Carolina Cretaceous white clay were used. The ice layers were thicker with depth below the surface (see Fig. 21) because cooling was slower and there was more time for the slow movement of water to the growing ice crystals. Carton No. 1 stood in sand saturated with water, while carton No. 2 was sealed at the bottom. The close similarity in the two tests proves that little water was drawn into carton No. 1, and this was confirmed by weighing the cartons before and after freezing.

FIG. 20.—Polygonal structure formed in bentonite on freezing. For description see text.

## EXPLANATION OF ICE SEGREGATION

The mechanics of the process by which an ice crystal grows under external pressure and displaces solids with which it is in contact is closely tied up with problems concerning the constitution of water and the nature of the changes occurring as the temperature is lowered and freezing takes place. Therefore, it is not possible to give a complete explanation at the present time, and only a tentative hypothesis is offered here.

During the growth of an ice layer in clay, water is supplied to the crystals through small capillary passages, but the upward flow of water should not be attributed to capillarity, as there is no free surface or meniscus. The uplift is due to the cohesive forces in the water. Askenasy's experiment shows that under suitable conditions a column of water may be placed under tension.[1] Conditions exist-

[1] Described on p. 134 of *Plant Physiology* by Vladimir I. Palladin. English edition edited by B. E. Livingston, Philadelphia, 1918.

ing during the growth of an ice layer in clay are particularly favorable for the uplift of water by molecular cohesion.

A growing ice crystal is in contact with a thin film of water similar to the adsorbed layer that forms on other solids. As molecules are removed from the film and attached to the crystal, they are replaced by others from the surrounding water. When an ice crystal grows in a direction in which growth is opposed by a solid body, such as a clay particle, the pressure is exerted through this thin film which

FIG. 21.—Cellular structure formed in bentonite on freezing. For description see text.

separates them. The film probably consists of little more than a single layer of molecules, for it is not expelled by pressure. After the available water has been exhausted, the film may be frozen, but it does not freeze easily. Cohesion is greater between the molecules in the film and between these molecules and the ice than it is between water molecules that are not close to ice crystals.

Orientation and attachment of a water molecule to an ice crystal is accompanied by a slight repulsion, proportional to the change in volume, and this results in a slight displacement of the whole crystal relative to the neighboring solid. Crystallization may be accompanied by the conversion of dihydrol into trihydrol molecules, as advocated by some physicists, but this would not materially alter

the process here outlined. As the new molecule is attached to the ice crystal, another molecule is drawn into the film by cohesion, and this is the direct cause of most of the displacement.

The growth of ice layers in soils and the accompanying pressure effects are therefore attributed to molecular cohesion. The energy for the process is of course supplied by the removal of heat. Water is not only pulled into the film under an ice crystal with force sufficient to lift the overlying load, but considerable force must be exerted in pulling it through dense clay to the growing crystal; hence in these experiments, water has been placed under tension sufficient to lift a column of water over 150 m. in height.

The maximum pressure that may be developed as a direct result of ice segregation in the materials used in these experiments is probably not much above 15 kg. per square centimeter. This limit is not due to the inability of ice crystals to grow under higher pressure, but to the failure of the water supply when the tensile strength of the water is exceeded. Water under such high tensile stress is superheated, even below 0° C., and the formation of vapor would immediately interrupt the supply. In one experiment water was forced into the clay under pressure, thus reducing the tensile stress at the points of crystal growth. This permitted segregation to take place under higher external pressure. When nitro-benzene, which boils at 210.85° C., was substituted for water, greater heaving was obtained under high pressure; but benzene, with a boiling point of 79.6°C., heaved less than water. Both of these liquids freeze with decrease in volume at temperatures between 5° and 6°.

## SUMMARY AND CONCLUSIONS

Freezing of soils often results in heaving much greater than that which can be attributed to change in volume of the water present. This excessive heaving is due to segregation of water as it freezes, additional water being drawn in by molecular cohesion. The pressure effects accompanying freezing result from the growth of ice crystals. Ice crystals develop pressure only in the direction of growth, which, in soils, is determined chiefly by the direction of heat conduction and the availability of water, differential pressure being of minor importance. In experiments a pressure of over 14 kg. per

square centimeter has been obtained through growth of crystals to form ice segregations, and water has been placed under tension sufficient to lift a column of water over 150 m. in length.

The chief factors controlling segregation and excessive heaving are: size of soil particle, amount of water available, size and percentage of voids, and rate of cooling. Segregation occurs readily if the particle diameter is less than a micron, and under favorable conditions where particles are somewhat larger. High water-content favors segregation, and additional water may be drawn from the water-table to form very thick ice layers. Water occupying very small soil voids does not freeze readily and may be undercooled in the immediate vicinity of ice crystals. Rapid cooling when the temperature is near the freezing point prevents segregation in some soils, but has little effect if the soil particles are sufficiently small.

Segregation of water on freezing causes shrinkage cracks if the supply of water is limited, or if the soil is very impermeable because of high colloid content. In the latter case, the cracks are gradually filled and enlarged by ice.

Frost heaving, when uniform, can cause little damage to pavements or structures, but unequal heaving may be very destructive. Local differences in soil texture, and in amount of available water, are the major causes of differential heaving. Differences in the kind and amount of soil cover are important, and difference in load is a minor factor.

Freezing, followed by thawing, may result in the concentration of so much water near the surface as to cause flowage and unequal settling. A sudden drop in temperature after a thaw which has left an excessive amount of water in the surface soil is a condition favoring segregation and excessive heaving.

The principles developed in this investigation help to explain certain geologic phenomena found in very cold regions, such as buried sheets, lenses and wedges of ice, and the soil polygons with convex surfaces. They also have many applications to engineering, especially in the construction of hard-surfaced roads. The resistance to freezing offered by water in very small pore spaces, and especially where colloids are present, may help to explain hardiness in plants.

# THE INDICES OF REFRACTION OF THE PLAGIOCLASE FELDSPARS

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ABSTRACT

A study of the accuracy of determining the composition of plagioclase feldspars by means of the available curves of indices of refraction demonstrates that consistent results are impossible. Two outstanding causes of this unfortunate situation seem to be: (1) inaccuracy of the old chemical analyses used as basis for the composition, and (2) the persistent assumption that the plagioclase feldspars consist of two and only two end members or minals.

Accurate data are needed before satisfactory curves can be drawn. The curves should be without excessive undulatory deviation from smooth ones, very close to straight lines. The composition should be in weight percentage.

## INTRODUCTION

For the last ten years I have become more and more convinced that the curves of the indices of refraction of the plagioclase feldspars used in determining the composition of these minerals in petrographic investigations give inconsistent results. For my own curiosity, I collected the curves which were readily available for comparative study. So pronounced are the dissimilarities between these curves that I feel this situation ought to be brought to the attention of mineralogists and petrographers with the earnest suggestion that fresh, modern, and accurate data be obtained as a basis for really satisfactory curves.

## GLOSSARY

There are two terms used in this paper that perhaps require a word in explanation:

1. *Minal*,[1] a natural compound, of definite chemical composition, possessing definite physical properties and frequently capable of assuming definite geometric forms, that with other minals forms (*a*) double salts, (*b*) crystalline solid solutions, and (*c*) mechanical intergrowths. In other words, they enter into the composition of natural minerals as actually found. Minerals, with a few exceptions, are composed of minals.

[1] H. L. Alling, *Jour. Geol.*, Vol. XXIX (1921), p. 218; Vol. XXXIV (1926), pp. 591–93.

TABLE I

LIST OF CURVES: INDICES OF REFRACTION; PLAGIOCLASE

| No. | Date | Source | Percentage | Form of Data | Basis | Reference |
|---|---|---|---|---|---|---|
| 1... | 1888 | Michel-Levy and Lacroix | Weight | Figures | Own | *Les Mineraux des Roches* (Paris), pp. 203–11 |
| 2... | 1894 | Fouqué | Weight | Figures | Own | "Contribution à l'étude des feldspaths des roches volcanique" *Bull. Soc. Min. Fr.*, Vol. XVII, pp. 283–611 |
| 3... | 1906 | Becke (see Fig. 1–1)* | Weight* | Curve and figures | Own | "Zur Physiographie der Gemeinteile der krystallinen Schiefer," *Denk. Akad. Wiss.*, Band LXXV (Wien), pp. 102–3 |
| 4... | 1909 | Winchell | Weight | Curve and figure | Compilation | *Optical Mineralogy*, p. 201 |
| 5... | 1909 | Larsen (see Figs. 1–6)* | Weight | Figure and curve $\frac{\alpha+\beta+\gamma}{3}$ | From literature | *Amer. Jour. Sci.*, Vol. XXVIII, p. 270, Fig. 3 |
| 6... | 1911 | Iddings (see Figs. 1–2)* | Weight | Curve | Compilation | *Rock Minerals*, p. 217 |
| 7... | 1913 | Wright (see Figs. 1–3)* | Molecular | Curve | Becke, Michel-Levy and Tertsch | A graphical plot for use in the microscopical determination of the plagioclase feldspars (*Amer. Jour. Sci.* [IV], Vol. XXXVI, p. 540) |
| 8... | 1917 | Bayley (see Figs. 1–4)* | Presumably weight | Figures | Compilation | *Descriptive Mineralogy* (Appleton), p. 421 |
| 9... | 1920 | Tsuboi (see Figs. 1–5)* | Presumably weight | Curve | Own | *Jour. Geol. Soc. Tokyo* (in Japanese), Vol. XXVII |
| 10... | 1920 | Niggli | Molecular | Curve | Iddings | *Lehrbuch der Mineralogie*, p. 245 |
| 11... | 1920 | Cotton and Peart | Weight | Curves | Iddings | *Jour. Roy. Soc. N.S.W.*, Vol. LIV, pp. 184–85 |
| 12... | 1922 | Johannsen | Molecular | Curve | Rosenbusch-Wülfing (Becke?) | *Essentials Microsc. Determ. Rock-forming Minerals and Rocks*, Fig. 24 |
| 13... | 1924 | Duparc and Reinhard | Weight | Figures | Becke and own | "La Determination des plagioclases dans coupe minces," *Mém. Soc. phy. et d'hist. nat., Genève*, Vol. XL, p. 134 |

TABLE I—*Continued*

| No. | Date | Source | Percentage | Form of Data | Basis | Reference |
|---|---|---|---|---|---|---|
| 14... | 1926 | Goranson (see Figs. 1–7) | Weight | Curve | Duparc, Reinhard and Becke | "The Determination of the Plagioclase Feldspars," *Amer. Min.*, Vol. XI, p. 152 |
| 15... | 1927 | Rosenbusch-Wülfing (see Figs. 1–8) | Molecular | Curve | Becke | *Mikroskoposke Physiographie*, Band I, Heft 2, p. 773 |
| 16... | 1927 | Winchell | Molecular | Curve | Wright and Tsuboi | *Optical Mineralogy*, Vol. II, p. 298 |

* Plotted in Fig. 1.

2. *Isontic*, a term proposed by Lane[1] "to apply to any line or surface drawn through points equal in some respect"; similar to isotherm, isobar, isocarbs, etc.; hence, "index isontics."

## THE CURVES OF INDICES OF REFRACTION

In Tables I and II, I have listed the data pertaining to the available curves. Undoubtedly there are others, but I feel that these are sufficient to show the need for fresh information. I have carefully plotted all of them on a uniform scale for my Lefax notebook. Believing that others would be interested, I have, in Figure 1, reproduced nine of them. Of course they are similar; perhaps some readers may be surprised that they are not identical. Some exhibit an undulatory character, while others show a fairly linear arrangement. Some curves for a given optical orientation are higher in value throughout the range in composition. Some of these differences are due to a difference in the unit in which the percentage of composition is expressed; the weight versus molecular per cent. The undulatory character is due in many cases to an apparently meticulous care in connecting all the points on the graph by continuous, constantly curving lines. This is illustrated by Goranson's No. 7 and Wright's No. 3 curves. Goranson's plot has been drawn from the data supplied by Duparc and Reinhard. I have taken the same data and constructed a set of straight-line curves, No. 9. I question whether the accuracy of the usual data is sufficiently high

[1] A. C. Lane, *Sci.* (N.S.), Vol. LXVIII (1928), p. 37.

to justify this meticulous treatment, and believe that its use results in less accurate determinations. We should not overrate the pre-

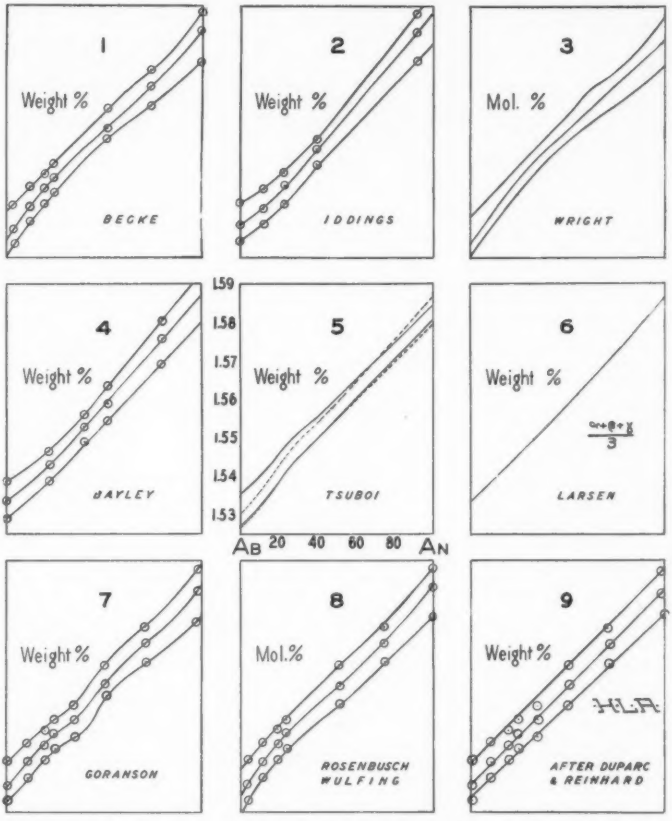


FIG. 1.—Reproductions, all on same scale, of curves of the indices of refraction of the plagioclase feldspars, revealing disconcerting differences and the paucity of data used in their construction. I believe there is no justification for the undulatory character of curves 3 and 7.

cision of the data. An Englishman would ask what is the "degree of tolerance" expressed by these curves. Wright says:

Their accuracy depends, therefore, directly on the data of observation. With the accumulation of more precise optical data on the plagioclases and

TABLE II

| No. | Name | Locality | Comp. as Given | | Date of Analysis | Comp. as Recast | | | Total Feldspar | Value of Indices | | | Used By | Composition Readjusted from 3 to a 2-Component System |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | Ab | An | | K-Feld. | Na-Feld. | Ca-Feld. | | α | β | γ | | |
| 1.... | Albite | Lakous, Crete | 100 | 0 | ? | ? | ? | ? | ? | 1.5290 | 1.5333 | 1.5386 | Iddings, Bayley | ..... |
| 2A.. | Albite | Amelia, Va. | 100 | 0 | 1882 | 1.72 | 93.83 | 4.45 | ..... | 1.5292 | 1.5331 | 1.5393 | Iddings | 96.5:3.5 |
| 2B.. | Albite | Amelia, Va. | 95 | 5 | 1882? | ..... | ..... | ..... | ..... | 1.5285 | 1.5321 | 1.5387 | Rosenbusch-Wülfing, Becke | 96.5:3.5 |
| 3.... | Albite | Greenland | 99.5 | 0.5 | 1908 | 2 | 98 | N.D. | 98.78 | 1.5283 | 1.5323 | 1.5386 | Duparc and Reinhard | Ab 100: An 0 |
| 4.... | Albite | Närestö | 100 | 0 | ? | ? | ? | ? | ? | 1.531 | 1.534 | 1.540 | Michel-Levy, Winchell, 1909 | ..... |
| 5A.. | Oligoclase | Bamle, Norway | 87 | 13 | 1869 | 2 | 84 | 14 | ..... | 1.534 | 1.538 | 1.542 | Michel-Levy | 87:13 |
| 5B.. | Oligoclase | Bamle, Norway | 87 | 13 | 1869 | 2 | 84 | 14 | ..... | 1.5341 | 1.5381 | 1.5431 | Duparc and Reinhard, Rosenbusch-Wülfing, Becke | 87:13 |
| 6.... | Ab-Olig. | ..... | 85.7 | 14.3 | ..... | ..... | ..... | ..... | ..... | 1.5386 | 1.5385 | 1.5433 | Winchell, 1909 | |
| 7.... | Oligoclase | Mexico | 85? | 15? | 1894 | 3.6 | 79.5 | 16.9 | 90.71 | 1.5358 | 1.5399 | 1.5436 | Fouqué | 85:15 |
| 8A.. | Oligoclase | Bakersville | 80 | 20 | 1888 | 3.3 | 74.5 | 22.2 | 98.07 | 1.5388 | 1.5428 | 1.5463 | Duparc and Reinhard, Rosenbusch-Wülfing, Becke | 79:21 |
| 8B.. | Oligoclase | Bakersville | 78 | 22 | 1888 | 3.3 | 74.5 | 22.2 | 98.07 | 1.539 | 1.543 | 1.547 | Michel-Levy | 79:21 |

TABLE II—*Continued*

| No. | Name | Locality | Comp. as Given | | Date of Analysis | Comp. as Recast | | | Total Feldspar | Value of Indices | | | Used By | Composition Readjusted from 3 to a 2-Component System |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | Ab | An | | K-Feld. | Na-Feld. | Ca-Feld. | | α | β | γ | | |
| 8C.. | Oligoclase | Bakersville | {78<br>80 | 22<br>20 | 1888 | 3.3 | 74.5 | 22.2 | 98.07 | 1.5389 | 1.5431 | 1.5469 | Iddings, Bayley, Winchell, 1909 | 79:21 |
| 8D.. | Oligoclase | Bakersville? | 80 | 20 | 1888? | 3.3? | 74.5? | 22.2? | 98.07? | 1.5385 | 1.5428 | 1.5465 | Winchell, 1927 | 79:21 |
| 9.... | Basic Olig. | Tvedestrand | 75 | 25 | 1845 | 7.45 | 69.62 | 22.93 | ..... | 1.5417 | 1.5458 | 1.5490 | Duparc and Reinhard, Rosenbusch-Wülfing, Winchell, 1909 | 80:20 |
| 10.... | Andesine | Hohenstein | 65 | 35 | ..... | 2 ?* | 61 ? | 37 ? | 98.92 | 1.5447 | 1.5489 | 1.5528 | Duparc and Reinhard | 62.5:37.5 |
| 11A.. | Andesine | Roche Sauve | 60 | 40 | 1883 | 4.4 | 48.4 | 47.2 | 95.51 | 1.549 | 1.553 | 1.556 | Michel-Levy, Fouqué, Iddings, Bayley, Winchell, 1909 | 53.5:46.5 |
| 11B.. | Andesine | Roche Sauve? | 60 | 40 | 1883? | 4.4? | 48.4? | 47.2? | 95.51? | 1.5500 | 1.5533 | 1.5570 | Winchell, 1927 | 53.5:46.5 |
| 12A.. | Labradorite | ..... | 50 | 50 | ..... | 4.4? | 48.4? | 47.2? | 95.51? | 1.5548 | 1.5578 | 1.5625 | Winchell, 1909 | 53:47 |
| 12B.. | Labradorite | ..... | 50 | 50 | ..... | 4.4? | 48.4? | 47.2? | 95.51? | 1.5547 | 1.5576 | 1.5622 | Winchell, 1927 | 53:47 |
| 12C.. | Labradorite | Coast of Labrador | 48 | 52 | 1878–88 | 3.18? | 43.81? | 53.01? | 100.54 | 1.554 | 1.557 | 1.562 | Michel-Levy | 53:47 |

TABLE II—*Continued*

| No. | Name | Locality | Comp. as Given | | Date of Analysis | Comp. as Recast | | | Total Feldspar | Value of Indices | | | Used By | Composition Readjusted from 3 to a 2-Component System |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | Ab | An | | K-Feld. | Na Feld. | Ca-Feld. | | α | β | γ | | |
| 12D.. | Labradorite | Coast of Labrador | 48 | 52 | 1878–88 | 3.18 | 43.81 | 53.01 | 100.54 | 1.5553 | 1.5582 | 1.5632 | Rosenbusch-Wülfing, Duparc and Reinhard, Becke | ..... |
| 13.... | Labradorite | Sta Lucia Pico | 48? | 52? | 1894 | 2.5 | 48.5 | 49.0 | ..... | 1.5545 | 1.5589 | 1.5634 | Fouqué, Bayley | 52:48 |
| 14.... | Labradorite | De la lava de 1870 Ile de Pico Açores | 48 | 52 | 1894 | 3 | 45 | 52 | 93.47 | 1.5556 | 1.5563 | 1.5631 | Fouqué | 49:51 |
| 15.... | Lab-Byt. | St. George Açores | 43? | 57? | 1894 | 0.7 | 42.5 | 56.8 | 99.54 | 1.5597 | 1.5628 | 1.5677 | Fouqué | 44:56 |
| 16.... | Labradorite | De Capello Faval | 43 | 57 | 1894 | 1.2<br>1.1 | 39.2<br>31.7 | 59.6<br>67.2 | 103.03<br>100.92 | 1.5611 | 1.5639 | 1.5689 | Fouqué, Winchell, 1909 | 33:67 |
| 17.... | Lab-Byt. | De la Beneyre Haute Loire | 40 | 60 | 1894 | 1.8 | 39.6 | 57.6 | 96.65 | 1.5617 | ..... | 1.5647 | Fouqué | 43:57 |
| 18A.. | Bytownite | Närødal | 27 | 73 | 1869 | 0?* 2. | 26<br>24 | 74<br>74 | ..... | 1.564 | 1.569 | 1.573 | Duparc and Reinhard | 26.5:73.5 |
| 18B.. | Bytownite | Närødal | 25 | 75 | 1869 | 0?* 2. | 26<br>24 | 74<br>74 | ..... | 1.564 | 1.569 | 1.573 | Rosenbusch-Wülfing, Becke | 26.5:73.5 |
| 18C.. | Bytownite | Närødal | 25 | 75 | 1869 | 0?* 2. | 26<br>24 | 74<br>74 | ..... | 1.5645 | 1.5693 | 1.5735 | Winchell, 1927 | 26.5:73.5 |
| 19.... | Bytownite | ? | 20 | 80 | ? | ? | ? | ? | ..... | 1.5691 | 1.5760 | 1.5805 | Bayley | ..... |

TABLE II—*Continued*

| No. | Name | Locality | Comp. as Given | | Date of Analysis | Comp. as Recast | | | Total Feldspar | Value of Indices | | | Used By | Composition Readjusted from 3 to a 2-Component System |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | Ab | An | | K-Feld. | Na-Feld. | Ca-Feld. | | α | β | γ | | |
| 20.... | Anorthite | Vesuvius | 9 | 91 | 1840 | 2.3 | 3.9 | 93.8 | ..... | 1.5752 | 1.5833 | 1.5884 | Bayley, Iddings | 7:93 |
| 21.... | Byt.-Anorth. | De St. Clement Puy-de-Dôme | 5 | 95 | 1890 | 0?* / 2. | 4.7 / 2.7 | 95.3 / 95.3 | 95.56 | 1.574 | 1.581 | 1.586 | Fouqué | 4.5:95.5 |
| 22.... | Anorthite | Vesuvius | 3 | 97 | ..... | 2 | 3 | 95 | ..... | 1.5748 | 1.5828 | 1.5880 | Duparc and Reinhard | ..... |
| 23.... | Anorthite | Vesuvius | 0 | 100 | ..... | ? | ? | ? | ? | 1.5756 | 1.5835 | 1.5885 | Rosenbusch-Wülfing, Becke | 5:95 |
| 24.... | Anorthite | Vesuvius? | 0 | 100 | ..... | ? | ? | ? | ? | 1.5755 | 1.5832 | 1.5885 | Winchell, 1927 | ..... |
| 25.... | Anorthite | ? | 0 | 100 | ..... | ? | ? | ? | ? | 1.5757 | 1.5837 | 1.5884 | Winchell, 1909 | ..... |

* $K_2O$ not determined. Assumed 2 per cent K-Feld. and subtracted it from Na-Feld.

especially with increase in knowledge of the effects of solid solution with orthoclase, carnegieite, nephelite, and possibly kaliophilite and other substances, these empirical curves will be changed somewhat.[1]

In a word, they are not very accurate, certainly not up to modern standards. Some of the causes for this rather discouraging state of affairs are discussed below.

## THE AGE AND RECASTING OF THE CHEMICAL ANALYSES

The most important matter concerning the chemical analyses the basis of all the curves under review, is their antiquity. Mineralogists and petrographers do not seem to realize that many of them are decidedly inferior. Washington[2] has been calling our attention to the real necessity of insisting upon high-grade, reliable analyses of rocks. In this respect the "Quantitative Classification of Igneous Rocks" (the "Norm" system) has been of inestimable value in raising standards. Insistence upon high-grade analyses of the common, rock-making minerals, however, is not so common. Most modern analyses are those of new minerals, for there is an unquenchable desire on the part of many mineralogists to find and describe new and hence rare minerals and neglect the common rock-making ones.

A glance at the column headed "Date of Analysis" in Table II will clearly show that the majority of the analyses are ancient. This fact alone introduces an uncertainty difficult to evaluate. Recasting is a method of testing that I have employed in every case,[3] when the analyses have been available to me. Of the fourteen analyses so tested only four reported the presence of $Fe_2O_3$, two of which noted a "trace." In one case even CaO was not determined! In none of the analyses was FeO and $H_2O$ given, and in a few cases $K_2O$ was determined by difference. MgO was reported only once! On recasting these analyses, the total feldspar was rarely close to 100 per cent, usually below. A range of 93–98 per cent was common.

[1] F. E. Wright, "A Graphic Plot for Use in the Microscopical Determination of the Plagioclase Feldspars," *Amer. Jour. Sci.* (4), Vol. XXXVI (1913), pp. 540–42.

[2] H. S. Washington, *U.S. Geol. Survey Prof. Paper 99* (1917).

[3] I have used Harry von Eckermann's "Molecular Quotienten," *Almquist and Wiksells Boktrycheri A.B.* (Upsala, 1925). These tables of molecular percentages, given to four decimals, are based upon atomic weights of 1911. I used a 20-inch slide rule for calculations not furnished by Eckermann.

Three analyses ranged 103–4 per cent, suggesting the presence of carnegieite-hephelite in solid solution.

Even with these results of recasting can we conclude that these particular analyses are very inferior? No, but I do believe that they are below present-day standards. I know, however, that modern up-to-date analyses, with some exceptions, on recasting, total close to a hundred, having a narrower range than those under discussion. In 1918 when I was recasting 1,300 analyses for plotting,[1] I was perplexed by the large number that gave incongruous results. On the assumption that all $K_2O$ was in $KAlSi_3O_8$, all $Na_2O$ in $NaAlSi_3O_8$, and all CaO in $CaAl_2Si_2O_8$, and calculating percentages with a slide rule, I found the total feldspar on addition differed considerably from the theoretical 100 per cent. I, perhaps unfortunately, was content to use 954 of these analyses even though the totals ranged from 85 to 110 per cent. Vogt[2] quite properly questions the wisdom of using many of these inferior analyses. Today I would reject a large number of them.

I have raised the fundamental question in using these recasts: What is the basis of recasting? Obviously that feldspars are polycomponent minerals composed of minals whose exact composition we have inferred from recasting analyses similar to those we are "testing"! If inconsistent results are obtained from recasting an analysis by F. W. Clarke,[3] as I have done, am I justified in rejecting the analysis even though it is modern and made by such a careful analyst just because our preconceived conceptions fail to fit? Or should we re-examine the basis for our recasting? If we attempt the latter, what shall we take as a starting-point? Surely fresh, new, careful, and complete analyses of specimens whose optical properties can be determined with high precision.

Granting that the analyses of the feldspars here discussed are inferior and incommensurate with modern demands, a further study of the results of recasting shows that the composition of these feld-

[1] H. L. Alling, *op. cit.*, Vol. XXIX (1921), pp. 254–55, Fig. 19.

[2] J. H. Vogt, "The Physical Chemistry of the Magmatic Differentiation of Igneous Rocks. On the Feldspar Diagram: Or:Ab:An," *Norske Videnskaps-Akademi* (Oslo, 1926), Vol. I, p. 51.

[3] "Oligoclase from Bakersville," *U.S. Geol. Survey Bull. 220* (1903), p. 28. Analysis made in 1888.

spars differs from that given in the "literature." Even though for the moment we ignore the presence of the potassic component, $KAlSi_3O_8$, the composition in simple ratios of Ab:An is frequently different from that given by the textbooks. These differences are often of such magnitude as to change the curves and to produce a corresponding change in the indices $\pm.0002$ to $\pm.0003$, hence the inconsistency in measuring indices to the fourth decimal, when a shift in composition by careful recasting will change the curve to that extent.

An illustration of the uncertainties that arise is that presented by the feldspar from Bakersville, North Carolina. This seems to be a favorite oligoclase. Michel-Levy, Fouqué, Iddings, Bayley, Duparc and Reinhard, Rosenbusch-Wülfing, and Becke list it. The composition is given as $Ab_{78}$ $An_{22}$ by Michel-Levy, Fouqué, Iddings and Bayley, while Duparc and Reinhard, Rosenbusch-Wülfing, and Becke give it as $Ab_{80}$ $An_{20}$. The indices are $\alpha$ 1.5388, $\beta$ 1.5428, and $\gamma$ 1.5463 by those who use $Ab_{80}$ $An_{20}$, and $\alpha$ 1.5389, $\beta$ 1.5431, and $\gamma$ 1.5469 by Fouqué, Iddings, and Bayley. Michel-Levy and Lacroix list $\alpha$ 1.539, $\beta$ 1.543, and $\gamma$ 1.547 as the indices. To add to this confusion there is the result of recasting the analysis by F. W. Clarke[1] of an oligoclase from Bakersville. The total feldspar is calculated to be 78.16. It is obvious that something is amiss. I can suggest that there are several old analyses of different specimens of feldspar from Bakersville and several determinations of indices have been made. Because of the difficulties encountered by the analysis by Clarke, I am compelled to use an old analysis, and hence am not sure I have made a proper correlation. Consequently, my calculated composition of K-Feld., 3.3; Na-Feld., 74.5; and Ca-Feld., 22.2 for this (or these) feldspar(s) from Bakersville is used with much uncertainty.

### THE PRESENCE OF THE POTASH MINAL IN PLAGIOCLASES

Another important matter concerning the chemical analyses is the consistent ignoring of the potash component $KAlSi_3O_8$, which is, in all probability, in solid solution in plagioclase feldspars. This is a matter upon which I have placed considerable emphasis during

[1] *U.S. Geol. Survey Bull. 220* (1903), p. 28.

the past seven years.[1] Duparc and Reinhard[2] do recognize that "practically, analyses frequently show the presence of $K_2O$ in small proportions in perfectly pure plagioclase. . . . . Nevertheless the influence of the potassic [minal] . . . . is too little known to allow us to take it into consideration and in practice we are restricted to consider albite and anorthite alone."[3] Goranson[4] says that "no attempt has been made to prepare curves for the determination of the amount of $KAlSi_3O_8$ in solid solution as the data available are not sufficient." This situation is regrettable and should be remedied speedily.

TABLE III

AVERAGE AMOUNTS OF $KAlSi_3O_8$ IN PLAGIOCLASE*

| | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| $NaAlSi_3O_8$.. | 0 | 10 | 20 | 30 | 40 | 50 | 60 | 70 | 80 | 90 |
| $CaAl_2Si_2O_8$. | 98.9 | 88.2 | 77.9 | 67 | 56.8 | 45.2 | 35 | 24.8 | 14.5 | 4.0 |
| $KAlSi_3O_8$... | 1.1 | 1.8 | 2.1 | 3.0 | 3.2 | 4.8 | 5 | 5.2 | 5.5† | 6.0† |

* Based upon J. H. Vogt, *op. cit.*, p. 48.
† Extrapolated by Alling.

The amounts of $KAlSi_3O_8$ in plagioclase is variable. Vogt[5] and I[6] have studied this matter and reached a partial solution of this problem, that is if 2 or 3 per cent of the potash component is assumed then greater accuracy is obtained. A more detailed statement is that with an increase in lime content there is a corresponding decrease in the potash minal. This is shown in Table III. If the potash feldspar minal is present in all plagioclase—and we certainly can safely assume it is in routine petrography—then any curves based upon chemical analyses which ignore $K_2O$ are defective and should be rejected. Some of the available data are in this respect unfortunate. If the potash minal is actually absent, then the data represent unusual feldspars which cannot be employed in setting up determinative curves for average specimens.

[1] H. L. Alling, *op. cit.*, Vol. XXIX (1921), p. 290; Vol. XXXI (1923), p. 358; Vol. XXXIV (1926), p. 593.

[2] *Mém. Soc. phy. et d'hist. nat., Genève* (1924), p. 4.

[3] Free translation.

[4] W. R. Goranson, *Amer. Min.*, Vol. XI (1926), p. 139.

[5] J. H. Vogt, *op. cit.*, p. 48.

[6] H. L. Alling, *op. cit.*, Vol. XXXI (1923), p. 364.

The absence of $K_2O$ in a chemical analysis, (usually indicated by a blank space), without notes to the effect "not determined," or "trace," is, to say the least, confusing. In this case I cannot assume that it was absent. The opposite attitude is taken by Duparc and Reinhard who state that they "have chosen whenever possible those types of plagioclase which do not contain potassium."[1]

If we could be assured that all of the curves under discussion represented potash-free feldspars, then we could use them with a constant correction with an assumed average percentage of $KAlSi_3O_8$. But we cannot make this assumption, for at least Duparc and Reinhard admit that in one case (that of basic oligoclase from Tvedestrand) the specimen contains 3 per cent of the potassic minal. Hence the curves are in reality projected ones, analogous to projected profiles in physiography or projected cross-sections in stratigraphy. The projected character is further brought out when the analyses are recast in terms of the three major minals and plotted upon a triangular co-ordinate basis as in Figures 2 and 3A. The necessity of recognizing the plagioclase feldspars as a three-component system is emphasized by this study.

But it is well to inquire how serious is the error when the potash minal is ignored and projected profiles used as though they were curves pertaining to a binary system. Figure 2 is instructive. It will be noticed that the double-ringed circles, indicating recast analyses, are, with a few exceptions, to the right and above the single-ringed circles. In other words, the composition of the plagioclase feldspars as calculated by recasting is different from the given composition in (1) containing the potash minal and in (2) a decrease in the soda component. I have elsewhere[2] suggested that more accurate results than customary would be obtained if 2 or 3 per cent of the potash minal is assumed outright, the curves being used to ascertain the lime component only; the per cent of the soda member to be secured by difference. The basis for this procedure is that, in all probability, the index isontics (lines of equal index)[3] of the tri-

[1] *Op. cit.* (translation), p. 5.

[2] H. L. Alling, "The Mineralography of the Feldspars. II," *op. cit.*, Vol. XXXI (1923), p. 361.

[3] A.C., Lane, *loc. cit.*

component system, K-, Na-, and Ca-feldspars, when plotted upon triangular co-ordinate paper, are approximately parallel to the potash-soda side of the triangle. The significant point is that the double-ringed circles of Figure 2 are, with some exceptions, to the right and above the single-ringed circles. In other words, they are located approximately where index isontics, if drawn,[1] would inter-

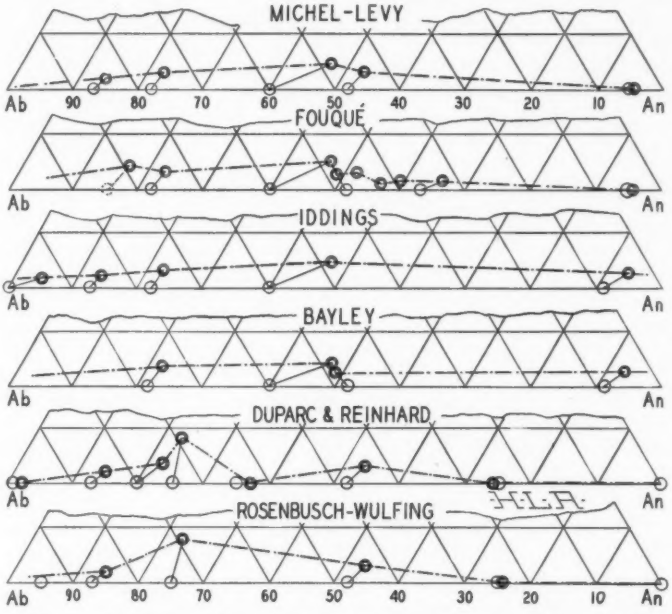


FIG. 2.—The lower portions of triangular plots of the composition of feldspars used by the above noted investigators in their attempts to establish curves of refractive indices. Single ring circles, composition as given; double ring circles, composition as found by recasting; heavy dot-dash line, the actual line of section of the curves.

sect circles of a corresponding pair. Hence, if the reader can follow this, the shift of the composition by recasting does not produce serious errors. Why is this so? The chemical separation of $Na_2O$ and $K_2O$ is a difficult one, and hence in analyzing plagioclase feldspars the temptation of the analyst is to assume all the alkali in

[1] H. L. Alling, "The Mineralography of the Feldspars. II," *op. cit.*, Vol. XXXI (1923), p. 361, Fig. 5.

them to be $Na_2O$, hence the feldspar in the opinion of a mineralogist is often considered potash free. In the analytical procedure the soda and the potash are weighed together and subsequently separated. In this manner some of this peculiar condition can perhaps be explained.

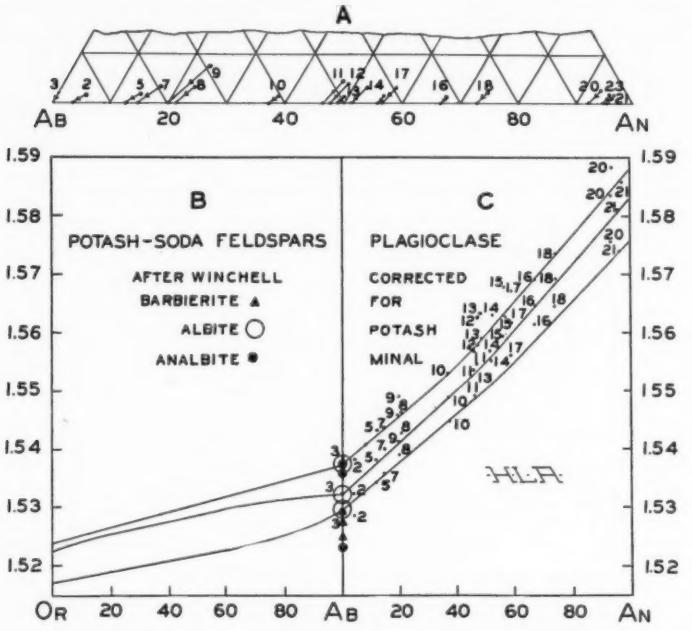


FIG. 3.—(A) The lower portion of a triangular plot of the composition of the feldspars listed in Table II as found by recasting the analyses in terms of three end members, K-, Na-, and Ca-Feldspars. This tricomponent system is reduced to a binary one by projecting the composition to the base, following the probable course of the index isontics. (B) A tentative and approximate plot of the indices of refraction of the orthoclase-albite system after Winchell and myself, and showing approximate index values of barbierite and analbite. (C) Index curves of the plagioclase feldspars, corrected for the potash minal, revealing the unsatisfactory character of the present data The use of this curve in routine work gives inconsistent results.

I have taken all of the data tabulated in Table II that provide us with chemical analyses which report $K_2O$, fourteen in number, and three others where I assumed 2 per cent potash minal, and plotted their composition in terms of the three common feldspathic

minals in Figure 3A. From a preliminary study of the index isontics[1] I have projected the composition back to the base of the triangle, as indicated by the arrows, thereby reducing it from a three- to a two-component system (see Table IV). The direction of the arrows was obtained from diagrams in the paper above referred to. Of course all of the curves under review here were offered as representing a binary one. But I would point out that the way to reduce the three-

TABLE IV

DATA OF TABLE II CORRECTED FOR POTASH MINAL
(SEE FIG. 3*a*)

| No. | % Ab | % An |
|---|---|---|
| 3 | 100 | .0 |
| 2 | 96.5 | 3.5 |
| 5 | 87 | 13 |
| 7 | 85 | 15 |
| 8 | 79 | 21 |
| 9 | 80 | 20 |
| 10* | 62.5 | 37.5 |
| 11 | 53.5 | 46.5 |
| 12 | 53 | 47 |
| 13 | 52 | 48 |
| 14 | 49 | 51 |
| 15 | 44 | 56 |
| 17 | 43 | 57 |
| 16 | 3.3 | 67 |
| 18* | 26.5 | 73.5 |
| 20 | 7 | 93 |
| 23 | 5 | 95 |
| 21* | 4.5 | 95.5 |

* Two per cent K-Feld. assumed and deduced from amount of Na-Feld.

component system to one of only two is not to ignore the potash minal, for inconsistencies are bound to occur. The method I have employed, I believe, provides us with more satisfactory data for the construction of these curves of refractive indices (see Fig. 3C). But it is well to realize that the curves in Figure 3C are not suitable for routine petrographic determination of the plagioclase feldspars. This set is corrected, as accurately as possible at the present time, for the potash minal, and inasmuch as natural plagioclases, as minerals (in contrast to the concept conveyed by the term minal), contain potash, these curves must be used by assuming outright that the third constituent, guided by Table III, is present.

[1] *Ibid.*

As Table II contains more data than any one set of curves under review, I transferred them to graph paper using the composition as given, and have reproduced it as Figure 4.

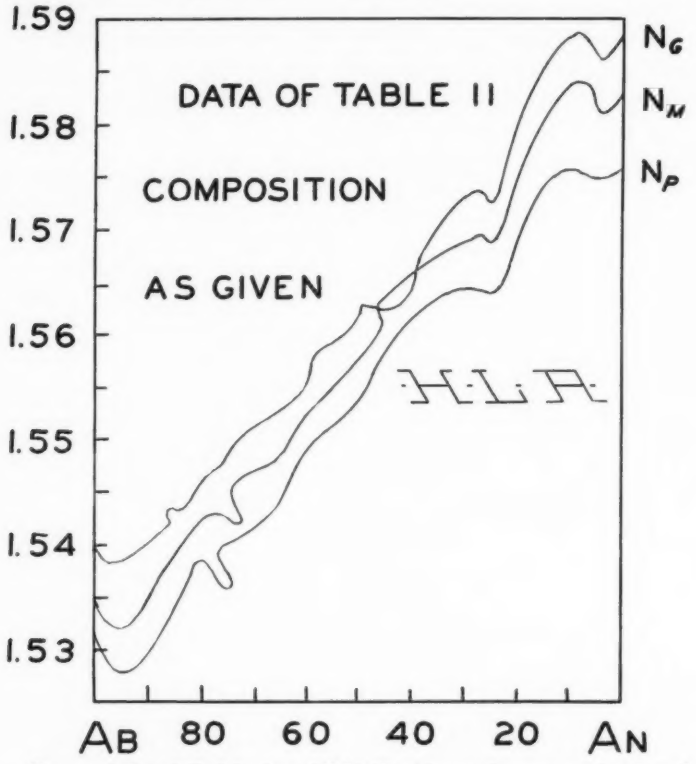


FIG. 4.—A plot of all of the data of Table II, with smooth curves passing through all the points, reproduced to show the unsatisfactory character of the data, various selections of which petrographers use.

It reveals a most disheartening situation. Smooth curves have been drawn through all the points as advocated by some mineralogists. It speaks for itself; it is most irregular and undulatory in the extreme. The reader may judge that too much liberty was exercised in drawing the smooth curves in Figure 3A, but he should look at Figure 4. And yet there are the data, various selections of which petrographers use!

### CAUSES OF DIFFERENCES BETWEEN CURVES

The curves reproduced in Figure 1 reveal individual differences. Why should they? Obviously they differ because different data were used. This statement is only partially true in that many mineralogists have employed much the same information. In most of them a number of feldspars are the same. It is obvious that considerable selection on the part of many individuals has been taken. It is also obvious that I cannot determine what was the basis for much of this selection; furthermore, it is possible to select a group of data, say only six, from the many that are available that will produce fairly regular curves. We all do this and think we are clever in obtaining such neat and satisfactory curves.

Now there are suggestions that $NaAlSi_3O_8$ not only occurs as albite but also as "barbierite" and "analbite." Winchell has presented evidence in support of such views,[1] and I have recently presented some consequences that follow if the soda minal is trimorphous.[2] Of course the idea of trimorphism does not meet with universal approval. Vogt[3] says he cannot agree, believing with Kalb[4] that the various forms are due to "highly varying physical and chemical conditions at the formation of the mineral." But Vogt[5] does not, in so far as I am able to discover, offer any evidence opposed to the concept of trimorphism.

I have considered the possibility that some of the causes of the differences between the curves of Figure 1 may be due to the presence of either barbierite or analbite. To test this I have drawn the curves for the refractive indices of the potash-soda feldspars $\beta$-series,[6] orthoclase-albite, as shown by Winchell,[7] and extended by

[1] A. N. Winchell, "Studies in the Feldspar Group," *Jour. Geol.*, Vol. XXXIII (1925), pp. 714–27.

[2] H. L. Alling, "The Potash-Soda Feldspars," *ibid.*, Vol. XXXIV (1926), pp. 591–611.

[3] J. H. L. Vogt, *op. cit.*, p. 13.

[4] *N. Jahrb. f. Min., etc., B.B.* (1924).

[5] *Op. cit.*

[6] H. L. Alling, "The Potash-Soda Feldspars," *op. cit.*, Vol. XXXIV (1926), pp. 591–611.

[7] A. N. Winchell, "Studies in the Feldspar Group," *Jour. Geol.*, Vol. XXXIII (1925), pp. 714–27.

me,[1] in Figure 3 B. The central line separating Figures 3 B and 3 C is soda feldspar, common to both the potash-soda and the plagioclase series. Upon this line I have indicated the approximate position of values of the indices ($N_p$, $N_m$, $N_g$, or $\alpha$, $\beta$, $\gamma$) of the barbierite and analbite. I reasoned (how correctly I don't know) that if the specimens of feldspar listed by the various investigators (Table II) contained any barbierite or analbite, the plotting of the refractive indices, as in Figure 3 C, would fail to lie on the curves connecting values for albite in the potash-soda series, Figure 3 B. A glance will show at once that a surprisingly satisfactory correlation is obtained, in that all values of the refractive indices of barbierite (triangles) are below those of albite (large open circles), and the $N_p$ ($\alpha$) value of analbite (small black circles) is below all three of barbierite. I am inclined to conclude, therefore, that polymorphous modifications of $NaAlSi_3O_8$ do not enter into the makeup of the plagioclase feldspars under review.

But inspection of Figure 3 C shows that feldspars 9, 12, 13, 14, 15, and 17 are reported as having values above the curves I have drawn. The last four feldspars were reported by Fouqué. Do these particular feldspars contain other constituents, such as carnegieite, kaliophylite,[2] celsian, the strontium and the potassium iron-feldspars, or possibly even the rare rubidium, lithium and ceasium feldspathic minals? A study of the recastings of the chemical analyses of these questioned plagioclases does certainly suggest that most of them are defective. Feldspars 9 and 12 total over a hundred and indicate carnegieite. Others, Nos. 14, 15, and 17, appear to be inferior in quality. But the suggestion of carnegieite does not explain the excessive values in index, for, on the contrary, Winchell[3] lists the indices as follows: $\alpha$ 1.509, $\beta$ 1.514, $\gamma$ 1.541—which are lower than any of the normal plagioclase minals. If we can assume, and here I am very uncertain, that the index isontics of the system, carnegieite-albite-anorthite, are curvilinear functions in spite of the lack of complete solubility between carnegieite and anorthite,[4] then the

[1] *Op. cit.*, p. 604, Fig. 3.

[2] See F. E. Wright, *op. cit.*, p. 540.

[3] A. N. Winchell, *The Optic and Microscopic Characters of Artificial Minerals* (Madison, 1927), pp. 141–42.

[4] N. L. Bowen, *Amer. Jour. Sci.* (1912), Vol. XXXIII, p. 551.

presence of carnegieite in solid solution in plagioclase would lower rather than raise the index.

Other minals that might be present which would raise the index are $KFeSi_3O_8$,[1] (which gives the gem "orthoclase" of Madagascar its pale-yellow color), celsian and strontium feldspars. As the analyses are so incomplete I have not adequate means of testing this suggestion. However, a simple calculation of the amount of $KFeSi_3O_8$ that would be necessary to account for the high value of the index of No. 13 shows that 9 per cent would be required. This is excessive. Similarly, 12 per cent celsian is needed. Hence I am at loss to account for these discrepancies unless the indices were (1) incorrectly determined, (2) determined by other than sodium light, (3) at dissimilar temperatures, (4) index determinations on one feldspar and the chemical analyses on another but supposed to be identical, (5) measured index represents more a calcic (core) of a zonal and the chemical analyses upon more sodic margins, and other factors I do not think of at the moment.

## CONCLUSIONS

1. There is a disheartening lack of uniformity between the various curves of the indices of refraction of the plagioclase feldspars used in microscopic determination (Fig. 1).

2. In attempting to assemble the data upon which these curves were drawn in order to ascertain the causes of these differences, unforeseen difficulties were encountered. Many of these difficulties were due to the lack of adequate references to the original sources. Even the more recent texts fail to assist the investigator. No writer, including myself, is above reproach in this matter.

3. The chemical analyses upon which these curves are based must be regarded as inferior. The method of recasting as a means of testing them is based upon the assumption that we possess sufficient knowledge of the end members of the feldspars. This may not be justified.

4. Too many of the curves are based upon unreliable data to meet modern requirements.

[1] H. L. Alling, "The Mineralography of the Feldspars. II," *op. cit.*, Vol. XXXI (1923), pp. 368–69; D. S. Belyankin and N. G. Sergiev, *Bull. Acad. Sci. union rep. soviet. social.* (VI), Vol. XX, pp. 1199–1206.

5. Too many of them are based upon too few specimens and the curves drawn with unjustified meticulous care in connecting all points by constantly curving lines. A good example of this unfortunate method is that of Goranson's (Fig. 1).

6. These curves are offered upon the assumption that they represent a binary system. This is no longer tenable. Potash feldspar is present in plagioclase. Hence these curves are in reality projected profiles (see Figs. 2 and 3A). Attempts to reduce this three-component system to one of two by ignoring the potash minal leads to discordant results. The method used in this study, and advocated, is of plotting the data as a three-component one and then projecting them to the albite-anorthite side (base) of the triangle, upon the basis of the probable course of the index isontics (Fig. 3A).

But a set of curves drawn upon a binary basis that is corrected for potash feldspar is not accurate in routine determinations because natural plagioclase contains potash feldspar. Even Figure 3C gives inconsistent results. I have a set of curves upon which I am working that recognizes the presence of the potash member but it is not yet ready for publication.

7. Additional constituents that enter into the actual composition of plagioclases are carnegieite, kaliophylite, celsian, strontium, lithium, rubidium, ceasium, and potash-iron feldspars. These effect the indices—some up and some down.

8. Uncertainties are prevalent, owing to difficulties in securing correct correlation between chemical analyses and optical data.

9. There is no assurance that the indices as reported are commensurate. Both the wave-length and the temperature must be known.

10. To sum up: The sets of curves of refractive indices of the plagioclase feldspars now available are not reliable and not up to present standards. New data, obtained under conditions of great precision, are needed.

# THE LATAH FORMATION IN IDAHO

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## ABSTRACT

It has been known for about thirty-five years that a fossil-bearing sedimentary series is interstratified with the Columbia River basalt throughout eastern Washington and northern Idaho. It has been observed and described in print by several geologists of the United States Geological Survey, as well as by workers on state surveys. In 1926, it was named the Latah formation in the Spokane area by Pardee and Bryan. Fossils collected by the writers from Russell's localities and from many new stations make certain that the Idaho localities represent exposures of the Latah series.

In Idaho, usually two or more layers occur separated by basalt flows. This offers persuasive evidence that the Latah is a series, rather than a formation. Although the lake beds are believed to be of local origin, and comparatively disconnected, the area in which outcrops occur is approximately 175 miles long and 75 miles wide.

The type locality, near Spokane, appears to be abnormal when its characteristics are compared with those of a large number of the widely distributed Idaho localities. This article is the result of only a preliminary study, and future work will no doubt yield new localities and more complete collections of the flora.

## INTRODUCTION

In 1923, Dr. Kirk Bryan named and defined the Latah formation in an address before a Columbia section meeting of the American Institute of Mining and Metallurgical Engineers. From the characteristics outlined it seemed obvious to some of the geologists present that he was describing a familiar formation, or series, which was known to occur in Idaho but which was unnamed. Certain features assigned to the formation at its Spokane locality were not characteristic of those in Idaho, and this investigation was undertaken to check these apparent differences.

Observations of the authors have, for the large part, been by-products of other geologic assignments, but many brief visits have been made to Latah localities for the specific study of the geologic relations, and for the collection of plant fossils.

## HISTORY OF THE LATAH FORMATION

Although Dr. Bryan named and defined the Latah formation in his address before the Institute members at Pullman, Washington,

November 10, 1923, it was not until January, 1926, that articles describing the formation[1] and flora[2] were published. Since that date the senior author has noted the occurrence of the Latah formation in seven different Idaho localities.[3] So far as known there are no other references in the literature to the Latah formation as such.

The occurrence of the lake beds, ash, sands, and silts interbedded with the Columbia River basalt has, however, been recognized by geologists for many years. Investigations and fossils collected by the present writers have shown that nearly all of these early papers described the series later to be named the Latah formation, by Pardee and Bryan.[4]

As early as 1893, Israel C. Russell[5] described leaf-bearing sediments which he referred to the John Day system, interbedded with the Columbia River basalt in central Washington. He says,

> This formation, as shown by its fossil leaves is of middle Tertiary (Miocene) age, and is here designated as the John Day system [p. 23].
>
> As already mentioned, there are thin beds belonging to the John Day system interstratified with the upper layers of Columbia lava and thin layers interbedded with the lower layer of the John Day system [p. 24].
>
> The quarry from which this rock was obtained at the south end of the Kittitas Valley contains clays and sandstones which are charged with fossil leaves. Other leaf bearing beds of the same character occur at John Day River.

[1] J. T. Pardee and Kirk Bryan, "Geology of the Latah Formation in Relation to the Lavas of the Columbia Plateau Near Spokane, Washington," *U.S. Geol. Survey Prof. Paper 140A* (1926).

[2] F. H. Knowlton, "Flora of the Latah Formation of Spokane, Washington, and Coeur d'Alene, Idaho," *U.S. Geol. Survey Prof. Paper 140A* (1926).

[3] Virgil R. D. Kirkham, "Ground Water for Municipal Supply at St. Maries, Idaho," *Idaho Bur. Mines & Geol. Pamphlet 17* (November, 1926); "Ground Water for Municipal Supply at Potlatch, Idaho," *Idaho Bur. Mines & Geol. Pamphlet 23* (June, 1927); "The General Geology of Eastern Washington and Northern Idaho," *Northwest Science*, Vol. I, No 2 (June, 1927); "Abstracts of Important Papers Dealing with the Geology of the Inland Empire and Adjacent Country," *Northwest Science*, Vol. I, No. 2 (June, 1927); "Underground Water Resources in the Vicinity of Orofino, Idaho," *Idaho Bur. Mines & Geol. Pamphlet 24* (July, 1927); "Underground Water Resources in the Vicinity of Lapwai, Idaho," *Idaho Bur. Mines & Geol. Pamphlet 24* (July, 1927); "Landslips Near Whitebird, Idaho," *Idaho Bur. Mines & Geol. Press Bull. 17* (July, 1928).

[4] *Op. cit.*

[5] Israel C. Russell, "A Geological Reconnaissance in Central Washington," *U.S. Geol. Survey Bull. 108* (1893).

It is by comparing the fossils from these two localities as well as by practically tracing the beds from one exposure to the other that the identity of the John Day system in general with the beds outcropping on John Day River has been established [pp. 23–24].

In a later article, Russell,[1] in describing sediments interbedded with basalt, says,

Owing to the presence of the sheets of clay and sand interleaved with the basalt in much of the region drained by the Palouse, springs are numerous along its branches . . . . [p. 27].

Interbedded with the sheets of basaltic rock composing the Columbia lava are beds of clay, sand, volcanic dust, and lapilli, which were deposited in part by streams and in part in lakes during intervals between lava flows [p. 29].

Above the Columbia lava in certain localities there are thick deposits of gravel, clay and volcanic dust, constituting what is termed the John Day system [p. 29].

From these statements it may be inferred that he had by this time recognized two or more distinct sedimentary series. On page 50 of the same article, Russell discussed the interstratified lake beds under the heading of "Clay and Gravel Interstratified with the Columbia Lava." Localities where interstratified lake beds and basalt occur are described on pages 50–55, and include the following headings: "Southeastern Washington," "Walla Walla Valley," "Asotin Country," "Palouse Country," and "Spokane." He found lake beds interstratified with the basalt at all of these localities, and made clear that these lake-bed sediments are not to be correlated with the John Day system, which he described in a separate section in the following pages of the report. He says,

Interleaved with the sheets of basalt forming the series named the Columbia lava, there are, as just shown, sheets of lacustral clay and of stream borne gravel and sand, and layers of volcanic dust and of volcanic fragments termed lapilli. The lake beds appear to occur mostly near the border of the lava-covered country, for the reason that, in Eastern Washington at least, the lava sheets flowed toward the mountains, and on congealing dammed their draining streams. . . . . The supposed absence of lake beds over the more central portions of the lava may be due to a lack of information, but seems to be a necessary result of the manner in which the lava sheets encroached on the mountains and blocked their draining streams. Shallow lakes were probably formed on

[1] Israel C. Russell, "Reconnaissance in Southeastern Washington," *U.S. Geol. Survey Water-Supply and Irrigation Paper 4* (1897).

the surfaces of the cooled lava sheets far from the mountains, but being at a distance from highlands that could furnish débris, they would not be deeply filled, and records of their existence would not be conspicuous [p. 55].

It is probable also that layers of clay and sand exist lower in the series than those penetrated at Pullman [p. 83].

Beds of clay and of fragments of volcanic rock (lapilli and volcanic dust) interbedded with basalt occur in the neighborhood of Asotin . . . . [p. 84].

During the intervals between the lava flows lakes were formed on the surface of the cooled basalt and sheets of sediments were laid down. . . . . Sand and gravel, swept out of the mountains by swift-flowing streams, were spread over the lava during intervals between the fiery floods. Alternating sheets of basalt, sand clay were thus formed, especially about the borders of the lava covered region [p. 90].

The next article in which Russell[1] described these interstratified lake beds and lavas was published in 1901. He says,

The lava was outpoured at successive intervals embracing a long period of time, as is shown by the occurrence, at several horizons, of layers of sedimentary material, principally clays and sand between the sheets [p. 29].

The greatest influence on the streams, however, occurred from the fact that the lava invaded the lowlands and entered the lower portions of the valleys in the mountains, thus obstructing the drainage and furnishing conditions for the origin of extensive lakes. The truth of this deduction is verified by the presence of lacustral deposits, some of them of great thickness and wide extent, interbedded with the lava sheets [p. 30].

The age of the lava, as determined principally from the fossil plants contained in the sedimentary layers and beds of volcanic dust interstratified with it, is Tertiary [p. 30].

On page 32 under the heading, "Sheets of Volcanic Dust," Russell gave a list of localities where they were observed, and says,

The evidence seems to indicate that in the region under review there are several layers of volcanic dust, some of them mingled with sedimentary material. . . . . These layers frequently contain fossil leaves, which afford excellent illustrations of the nature of the varied and luxuriant flora which clothed the lava plain at the time the dust showers occurred.

On page 35 under the heading, "Sedimentary Beds," Russell described at considerable length lake beds interstratified with the basalt. And on pages 35–42, he described well logs and sections in many Idaho localities. On page 41 he says, "From such facts as are

[1] Israel C. Russell, "Geology and Water Resources of Nez Perce County, Idaho," *U.S. Geol. Survey Water-Supply Papers 53 and 54* (1901).

in hand it is evident that in the Nez Perce region, where extensive lava plateaus occur, the presence of porous water-bearing beds is to be expected at three principal horizons, namely, at 150 to 350 feet, at 750 to 1000 feet, and at 3,500 to 3,600 feet." In associated pages he made clear that these water-bearing beds are lake beds overlain, separated, and underlain by Columbia River basalt.

The interstratified lake beds and lavas were next described in 1904, in an article by Lindgren,[1] who found the same conditions to exist as did Russell. He says,

> In a few places waterlaid clays and sands, with local lignites, are intercalated, doubtless accumulated by the rivers during lulls in the eruptions [p. 24].
>
> The age of the basalt is generally conceded to be Miocene. The evidence, consisting mostly of fossil plants in interbedded sediments, has been gathered from various points and examined, chiefly by Professors Knowlton and Merriam [p. 24].
>
> Intercalated between the basalt flows are occasionally found thin beds of white volcanic dust of rhyolitic character [p. 76].
>
> Waterlaid beds, consisting of clays and sands swept down from the surrounding mountains, occur in some places between the lava flows [p. 76].

Landes,[2] in an article published in 1905, agreed with Russell in his *Bulletin 108* and *Water-Supply Paper 4*. He says, in referring to the deep artesian wells at Pullman, Palouse, and Uniontown in Eastern Washington, "These wells enter the rock and obtain water from beds of sand and gravel intercalated with basalt [p. 59]."

Lupton[3] described sedimentary beds interstratified with the basalt in the vicinity of Orofino, Idaho, in an article published in 1915. He accepted Russell's findings regarding the relations of these lake beds to the lava.

In December, 1923, Laney, Kirkham, and Piper[4] described sediments interstratified with basalt in Idaho.

[1] Waldemar Lindgren, "A Geological Reconnaissance across the Bitter Root Range and Clearwater Mountains in Montana and Idaho," *U.S. Geol. Survey Prof. Paper 27* (1904).

[2] Henry Landes, "Preliminary Report on the Underground Waters of Washington," *U.S. Geol. Survey Water-Supply Paper 111* (1905).

[3] Charles T. Lupton, "The Orofino Coal Field, Clearwater, Lewis, and Idaho Counties, Idaho," *U.S. Geol. Survey Bull. 621* (1915).

[4] F. B. Laney, V. R. D. Kirkham, and A. M. Piper, "Ground Water Supply at Moscow, Idaho," *Idaho Bur. Mines & Geol. Pamphlet 8* (December, 1923).

When Bryan announced a name for the aforementioned series in 1923, it seemed obvious to the senior author that it was the same series described in the last seven papers referred to above. The Latah formation at Spokane, by definition, occupies a position underlying the Columbia River basalt instead of being interstratified with the lava. This notable difference, along with several others claimed for the Spokane locality, prompted the writers to visit approximately 40 localities in Idaho, as well as the Spokane localities listed by Pardee and Bryan. The majority of Russell's localities were visited, and new ones were discovered. From nearly all of these, plant fossils were collected. Several lots were sent to the United States Geological Survey for determination, and some were determined by paleobotanists of other organizations. Nearly all of the specimens submitted from each locality were identified as typical of the Latah flora.

In order that the significance of the data in the following pages may be more clearly understood the reader is invited to study the following summary, presented by Pardee and Bryan[1] on the first page of their article.

> The name Latah formation is proposed for a series of beds consisting mostly of clay and shale and of fresh-water origin that are found near Spokane, Wash., and that contain an abundant middle or lower Miocene flora. The formation is found in Latah Creek Valley for 10 miles south of Spokane, on tributaries of Little Spokane River for 20 miles north of the city, and in places in Spokane Valley for 8 miles northwest and 5 miles east of the city. It occurs also in at least one place in the broad, open valleys of the mountains near Coeur d'Alene, Idaho. These beds rest upon a rough surface of granitic and schistose rocks of unknown age, attain a maximum thickness of 1,500 feet, and are overlain by "rim rock" basalt flows, probably Miocene; "valley" flows, probably later Tertiary; and gravel of Pleistocene age. Sills and dikes, which represent some of the conduits through which the "rim rock" basalt rose to the surface, cut and to a small extent deform the Latah beds, which otherwise retain their original horizontal attitude.
>
> During Latah time the lava floods of the Columbia Plateau, advancing from the west, were held back from the Spokane area by a ridge, the highest parts of which now form the hills near Cheney, Marshall, and Medical Lake. The drainage outlet of this area was thus gradually obstructed by the lava, and the sediments here described were accumulated. Eventually the lava piled up until the "rim rock" flows were enabled to cross the protecting ridge and to overspread the area of Latah sediments.

[1] *Op. cit.*

The "rim rock" flows consist of olivine basalt that shows very conspicuous columnar jointing. They are considered to be of Miocene age but are doubtless younger than the Yakima basalt. West of the submerged ridge they are apparently indistinguishable from the great series of flows that make up the Columbia Plateau.

After a time interval measured by the cutting of valleys through the "rim rock" basalt and to a known depth of 800 feet into the Latah formation another series of basalt flows, referred to as "valley" flows, were extruded. These flows occupy parts of the erosional valleys and rest unconformably upon the Latah, exhibiting at the contact "pillow structure" and other characteristic features. They are distinguished from the "rim rock" flows largely by their topographic position, by a more varied jointing, and by the absence of olivine.

## LOCAL DETAILS (SEE FIG. 1)

The following notes for various stations in Idaho give briefly the location, relations, and other essential data for each exposure. The numbers of the stations are for convenience only and do not represent chronological order of study. The accompanying key map indicates their location.

*Station 4* is about 11 miles east of Lewiston, Idaho, on the north bank of the Clearwater River, near the highway in T. 36 N., R. 4 W., B. M. The Latah beds are composed chiefly of ash and white shale and gradations of both. These beds dip southwest at an average angle of 20 degrees. A collection made here included 12 species of plants.

*Stations 1, 2, and 3* lie west of this station within a distance of half a mile. Station 1 shows the Latah lying conformably under Columbia River basalt. A collection contains 9 species of plants.

*Station 5* is a few yards east of Station 4. The Latah lies conformably on the Columbia River basalt, dips southwest and west at angles between 15 and 20 degrees. Six species of fossil plants were collected here. Several hundred feet of basalt overlie the Latah at Station 1, and a similar thickness of basalt underlies it at Station 5. About 350 feet of Latah formation lie between the basalt flows at this place.

*Station 6* is in Sec. 6, T. 36 N., R. 2 E., B. M., about 1 mile northeast of Orofino, Idaho, on the east bank of the Clearwater River. Here the Latah beds consist of shale, sandy shale, and ash. An apparent thickness of approximately 200 feet of Latah lies between two layers of Columbia River basalt, each of which is about 300 feet thick. The basalt and Latah series dip gently westward at less than 2 degrees. A collection of fossils was made from very fragile shale which contained specimens identified in the field as probably *Populus* and *Sequoia*. Nothing was determinable after shipment.

*Station 7* is in Sec. 15, T. 35 N., R. 4 E., B. M., at Weippe, Idaho. A well log and cuttings from a drilling well revealed 30 feet of sand, shale, and charcoal

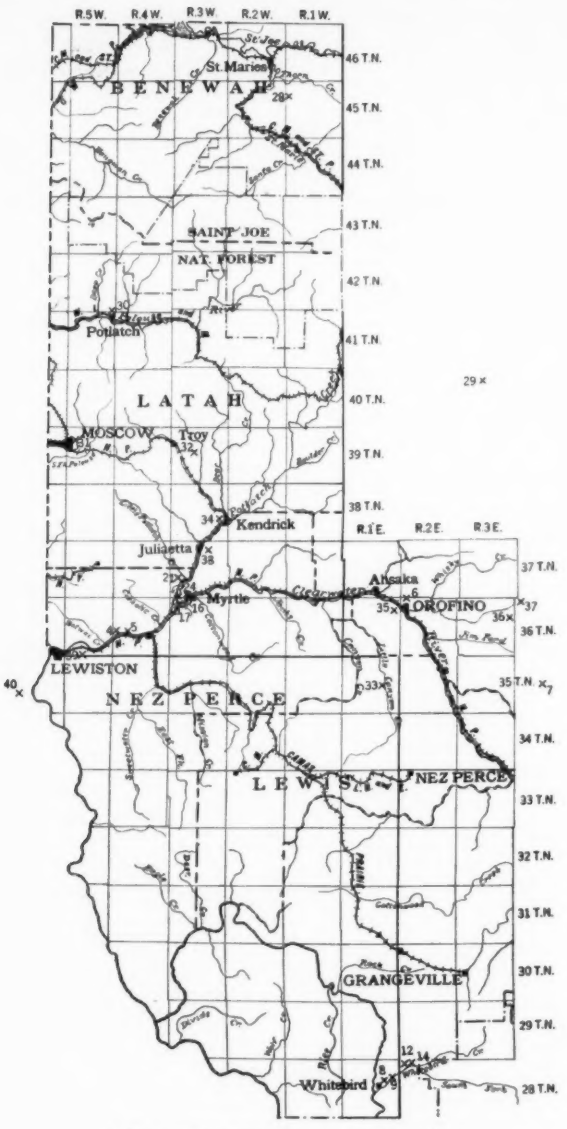


FIG. 1.—Key map showing Latah localities

underlying 30 feet of basalt. The well had not passed through the sedimentary series, but nearby wells with similar logs, drilled through 50 to 65 feet of lava, 75 feet of shale, clay, and charcoal, and into underlying basalt. Much coniferous charcoal and fragmental shale containing leaf impressions were found in the cuttings. A layer of charcoal 7 feet thick was passed through in all the wells.

*Station 8* is less than a mile east of Whitebird, Idaho, on the north side of Whitebird Creek, on the North-South Highway. The beds are chiefly ash. The exposure is along the strike, which is east and west. Dips are about 40 degrees to the north. A thickness of about 12 feet is exposed, with overburden of gravel. A collection of poor fossil material was made in which only 1 species was finally identified.

*Station 9* is slightly more than a mile east of Whitebird, Idaho, on the north side of Whitebird Creek, on the North-South Highway. The beds are chiefly ash and shale. They show a thickness here of 25 feet, strike east and west, and dip north at nearly 40 degrees. A small collection of fossils revealed 17 species of plants, of which all but 2 are found at the Spokane localities.

*Station 12* is 3 miles east of Whitebird, Idaho, on the north side of Whitebird Creek, on the North-South Highway. The beds are chiefly ash. A thickness of about 15 feet is exposed along the strike, with gravel overburden. The dip is obscure, but is apparently to the north. A small collection of poor fossil material contained only 3 determinable species.

*Station 14* is about 3.8 miles east of Whitebird, Idaho, on the north side of the road where it turns northwest and leaves Whitebird Creek. Several hundred feet of shales and ash beds are well exposed by the road cutting across the strike. The beds strike east and west, and dip to the north at angles ranging from 15 to 25 degrees. A collection of fossils made here contained 36 species of plants, nearly all of which are found at Spokane.

*Stations 10, 11, and 13* show outcrops of the Latah dipping northward. No collections were made. The entire series lies under at least 800 feet of basalt, and above basalt of an unknown thickness. Both basalt series dip north at comparable angles.

*Station 16* is 1.1 miles east of Arrow Junction, Idaho, on the east bank of Clearwater River, and east of the highway. Between flows of Columbia River basalt lie 12 feet of Latah shales, about 3 feet of which are fossiliferous. The beds strike N. 85 degrees E., and dip 25 degrees south. A small collection of poor fossil material contained 11 species of plants.

*Station 17* is slightly less than a mile east of Arrow Junction, Idaho, on the east bank of the Clearwater River, and near the highway. More than 50 feet of ash and shales are exposed between flows of Columbia River basalt. The beds strike N. 80 E., and dip 22 degrees SE. A small collection of poor fossil material contained 10 species of plants.

*Station 24* is about half a mile northeast of Arrow Junction, Idaho, on the east bank of Potlatch Creek. About 200 feet of Latah shales are exposed lying between flows of Columbia River basalt. The best fossil horizon lies a few feet below the upper basalt series. The beds strike N. 15 degrees E., and dip steeply

to the southeast. A small collection of fossil material contained 7 species of plants.

*Stations 18 to 23* show obscure outcrops of shale where collections were poor and contacts of the Latah and basalt were not clear.

*Station 27* is about 5 miles south of Julietta, Idaho, at Albright Hill, on the west side of the road. About 15 feet of Latah shales are exposed between flows of Columbia River basalt, dipping to the southwest at an angle of less than 2 degrees. The top of the Latah is metamorphosed, and the fossiliferous horizon lies from 4 to 6 feet above the road bed. About 900 feet of basalt overlie the Latah beds here. A collection of fossil material from this locality shows 26 species of plants.

*Stations 25 and 26*, respectively, are about 1 mile, and 1½ miles north of Arrow Junction, on the west side of Potlatch Creek, where the Latah formation is exposed. A collection made at Station 26 contains 5 species of plants.

*Station 28* is near St. Maries, Idaho. The following quotation from Kirkham's article,[1] of 1926, describes this occurrence,

"This formation, made up of silts and sediments, deposited in temporary lakes and ponds, is exposed in a limited outcrop at only one place in the area as far as could be observed during this brief examination. This exposure is in a cut in basalt on the Santa road in the extreme southeastern corner of the region. In Sec. 7, T. 45 N., R. 1 W., these silts, scarcely consolidated, were exposed for a length of perhaps 150 feet and for a thickness not in excess of 100 feet between two lava flows of the above mentioned Miocene basalt. Their age is thus established as being contemporaneous with the basaltic extrusions and they likewise give some criteria by which to judge the time interval which must have elapsed between these two flows. The outcrop lies outside the limits of the mapped area. Carbonaceous and lignitic seams indicate the presence during deposition of swampy growth and plant material. For the most part, the material represented mud, clay, and silt in strata lying approximately horizontally.

"From the field relations, as well as from the fact that the formation was not observed to be exposed in any other part of the region, together with the absence of notable terraces in the basalt flows themselves, persuasive evidence is offered that these sediments were local in character and confined to the lower part of the stream valley. They appear to be between the first and second flows and were probably deposited in small lakes or ponds formed as a result of changes in drainage caused by the original lava flow. As the damming was relieved, the ponds and lakes were eliminated and much, and perhaps nearly all, of the unconsolidated sediments were washed away and cleared from the surface of the bottom flow and valley walls. Naturally, occasional erosional remnants were left which were enclosed by the second flow and only revealed by haphazard erosion or by the work of man which, in this case, revealed this remnant by building a highway. No widespread deposit of this nature is believed to lie between any of the basalt flows exposed in the mapped area, al-

[1] *Op. cit.*

though varying thicknesses of soil, up to a few feet, undoubtedly separate nearly all of the flows. Water was seeping into the road from these silts and clays, where exposed, although no spring emerges from them."

Plant fossils, including *Populus* were noted here but no collections were made for identification.

*Station 29* is in T. 40 N., R. 3 E., about 8 miles north of Elk River, Idaho. Sediments consisting of flat-lying ash and shales, and containing *Taxodium*, underlie Columbia River basalt. The bottom of the sediments is not exposed.

*Station 30* is at Potlatch, Idaho. The following quotation from Kirkham's[1] article describes this occurrence,

"Interbedded with basalt flows in this area occurs a formation commonly found associated with the basalt along the eastern and southeastern edges of the Columbia River lava plateau. It has been called the Latah formation at Spokane by Pardee and Bryan of the U.S. Geological Survey. It has also been described at St. Maries.

"As shown by well logs at Potlatch, this formation here shows the following members from bottom to top starting at a depth of 566 feet.

| | | |
|---|---|---|
| Sandstone | 31 | feet |
| Coarse conglomerate | 3 | " |
| Alternate layers of red and brown clay | 57 | " |
| Basalt | 28 | " |
| Shale | 18 | " |
| Conglomerate | 21 | " |
| Clay | 8 | " |
| Loosely cemented sandstone | 14 | " |
| Well cemented sandstone | 15 | " |
| Basalt | 6 | " Bottom 566 feet. |

"Above this occur four lava flows of 91, 66, 78, and 70 feet thickness, respectively. On these lie 47 feet of blue clay which is covered by 19 feet of soil, gravel, and surficial overburden.

"The total thickness of sediments represented here, exclusive of basalt flows and surficial overburden, is 215 feet. Much of this material serves as an excellent aquifer."

Leaf fragments from well cuttings were common but none were determinable.

*Station 31* is at Moscow, Idaho. A deep well drilled in 1924–25, to underlying granite and quartzite, passed through 36 feet of basalt, a thin layer of sand and gravel, another basalt flow of 82 feet, a thin layer of shale, 65 feet of basalt, and a layer of clays, sands, and shales, 310 feet thick, lying on quartzite. The conditions here are essentially the same as those described by Russell[2] at Pullman. Much lignitic wood and charcoal were found in the well cuttings, but so far as is known, no determinable leaf fragments were found. Although

[1] *Op. cit.* [2] *Op. cit.*

2 layers of sand, clay, and gravel were found interstratified with the lavas, most of the series underlies the lowermost basalt flow. In another Moscow well 98 feet of sand and clay overlie 207 feet of basalt, which is underlain by 25 feet of sand and clay.

*Station 32* is in T. 39 N., R. 3 W., in the eastern part of Latah County, Idaho, along Bear Creek, about from 4 to 9 miles east of Troy, Idaho. Here a widespread bed of clay and shale underlies the upper basalt layers and occurs above the lower layers. It is exposed in the valley walls and in a large number of wells. Plant fossils were found in it at a number of places but none were submitted for determination. An area of at least 12 square miles of this layer is attested by wells and outcrops. The overlying lava represents from 2 to 5 flows. At places 2, and perhaps 3, distinct sedimentary layers are exposed, all of which lie between basalt flows.

*Station 33* is in T. 35 N., R. 1 E., about Sec. 14, approximately 9 miles south of Orofino, Idaho. This locality is mentioned by Russell[1] and Lupton.[2] It was visited by the writers in June, 1927, and showed shales and silts interbedded with lavas. The layer of sediments cropped out at about 800 feet below the top of the lava plateau. An unmeasured thickness of basalt underlies it. Considerable plant and carbonaceous material was present but no collections were made.

*Station 34* is in T. 38 N., R. 3 W., about 3½ miles northwest of Kendrick, Idaho, on the grade of the Kendrick-Troy road. The brief investigation showed ash, shale, and diatomaceous earth containing leaf fossils, beneath about 250 feet of basalt. Several hundred feet of basalt underlie the sediments.

*Station 35* is in T. 36 N., R. 1 E., and R. 2 E., northeast of Orofino, on both sides of the Clearwater River. Several localities included here are mentioned by Russell[3] and Lupton.[4] Nearly all exposures were in slide material. At no locality was it certain that the material was in place. The beds were interstratified with basalt at every place observed, but in the slide material this may, or may not, have significance. The layer is thought to be about 200 feet thick and is believed to lie below about 750 feet of basalt. Leaves of *Populus*, *Sequoia*, and *Taxodium* were found, but no collections were determined because of the uncertainty of the position of the material.

*Station 36* is in Sec. 12, T. 36 N., R. 3 E., in the bed of Orofino Creek about 15 miles east of Orofino, Idaho. An old coal prospect mentioned by Russell,[5] and Lupton,[6] is now inaccessible. Only float made up of ash and shale was visible without stripping. Talus from several hundred feet of overlying lava has greatly obscured the outcrop. Leaf fossils were present in some of the shale float. According to Russell,[7] the sedimentary bed lies in the bottom of the creek, and whether or not lava lies under it is not known. The horizon appears

[1] *Op. cit.*
[2] *Op. cit.*
[3] *Op. cit.*
[4] *Op. cit.*
[5] *Op. cit.*
[6] *Op. cit.*
[7] *Op. cit.*

to be the same one that occurs at other places in this region, as the lava canyon is about 800 feet deep here.

*Station 37* is in Sec. 6, T. 36 N., R. 4 E., about a mile east of Station 36. Conditions here are essentially the same as at Station 36.

*Station 38* is in Secs. 34, 35, and 36, T. 38 N., R. 3 W., from about 3 to 5 miles east of Julietta, along the road grade. This locality is mentioned by Russell,[1] and represents a bed of ash, shale, and sand in a place below about 250 feet of basalt. Other outcrops are thought to be slide material. No collections were made here because the material was very poorly consolidated.

*Station 39* is near the Ira Small well, 6 miles east of Lewiston, Idaho. Russell[2] mentions a well here as showing 160 feet of sand and clay overlain by 220 feet of basalt, with at least 75 feet of basalt underlying the sediments.

*Station 40* is in Asotin Canyon. In this locality Russell[3] gives a section of 155 feet of solid and broken basalt, overlying 110 feet of sedimentary beds, which is in turn underlain by 150 feet of basalt. This overlies 140 feet of sediments, which are underlain by at least 390 feet of basalt.

Of the 40 localities mentioned in the foregoing, Stations 1 to 5, 24 to 27, 33, 35, and 38 to 40 represent localities described in Russell's publications. Stations 33, 35, 36, and 37 represent localities mentioned by Lupton.[4] The others are described here for the first time or as cited. Many more new localities are likely to be found when more detailed work is done in the area.

A collection of fossils from 13 of these stations proves the sediments under discussion to be members of the Latah formation or series (see Table I). Complete collections were not made at any station, and only sufficient material was taken to establish the fact that the sediments actually represent the Latah formation. At the Spokane locality collections have been made over a period of years by many investigators and an attempt to fill out a complete suite has yielded a long list of species. The largest number of species yielded by any Idaho locality described here is 36, although another locality yielded 26. The others gave 12 species or less.

Careful collecting from Stations 1, 4, 9, 14, and 27 will very likely augment the species known from those localities until the list may even rival the Spokane locality. Table I shows the species known at the Spokane localities and at 13 Idaho localities, exclusive of Coeur d'Alene.

[1] *Op. cit.*
[2] *Op. cit.*
[3] *Op. cit.*
[4] *Op. cit.*

TABLE I

| | Spokane Localities | Idaho Localities | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | 1 | 4 | 5 | 7 | 8 | 9 | 12 | 14 | 16 | 17 | 24 | 26 | 27 |
| Acer bendirei Lesquereux | × | | | | | | | | | | | | | |
| Acer chaneyi Knowlton | × | | | | | | | | | | | | | × |
| Acer florissanti Kirchner | | | | | | | | | × | | | | | × |
| Acer merriami Knowlton | × | × | | | | | | | | × | | × | | × |
| Acer minor Knowlton | × | × | | | | | | | | | | | | |
| Acer oregonianum Knowlton | | | | | | | | | × | | | | | |
| Acer vivarium Knowlton | | | | | | | × | | | | | | | |
| Acer—not named | | | | | | | × | | × | | | | | |
| Acorn cup | | | | | | | × | | | | | | | |
| Alnus cones | × | | | | | | | | | | | | | |
| Amelanchier grayi Chaney | | × | | | | | | | | | | | | |
| Archaeomnium patens E. G. Britton | × | | | | | | | | | | | | | |
| Arisaema hesperia Knowlton | × | | | | | | | | | | | | | |
| Betula bryani Knowlton | × | | | | | | | | | | | | | |
| Betula fairii Knowlton | × | | | | | | | | | × | × | | | |
| Betula heteromorpha Knowlton | × | | | | | | × | | × | | | | | |
| Betula? largei Knowlton | × | | × | | | | | | | | | | | |
| Betula nanoides Knowlton | × | | | | | | | | | | | | | |
| Betula thor Knowlton | × | | | | | | | | | × | × | × | | |
| Branchlet and bud? | × | | | | | | | | | | | | | |
| Calyx, cf. Porana | | | | | | | | | × | | | | | |
| Carpinus sp | × | | | | | | | | | | | | | |
| Carpites boraginoides Knowlton | × | | | | | | | | | | | | | × |
| Carpites ginkgoides Knowlton | × | | | | | | | | | | | | | |
| Carpites magnifica Knowlton | × | | | | | | | | | | | | | |
| Carpites menthoides Knowlton | × | | | | | | | | × | | | | | |
| Carpites paulownia Knowlton | × | | | | | | | | | | | | | |
| Carpites polygonoides Knowlton | × | | | | | | | | | | | | | |
| Carpites spokanensis Knowlton | × | | | | | | | | | | | | | |
| Carpolithus sp | | | × | | | | | | | | | | | |
| Cassia idahoensis Knowlton | | | | | | | | | × | | | | | |
| Cassia sophoroides (Kn) Berry | | | | | | | × | | | | | | | × |
| Castanea castaneaefolia (Unger) Kn | × | | | × | | | | × | × | | | | | |
| Celastrus fernquisti Knowlton | × | | | | | | | | | | | | | |
| Cercis? spokanensis Knowlton | × | | | | | | | | | | | | | |
| Cercis idahoensis n.sp. leaves and pods | | | | | | | | | × | | | | | |
| Charcoal, coniferous | | | | | × | | | | | | | | | |
| Diospyros andersonae Knowlton | × | | | | | | | | | | | | | |
| Diospyros? microcalyx Knowlton | × | | | | | | | | | | | | | |
| Equisetum, underground stem | × | | | | | | | | | | | | | |
| Euonymus knowltoni Berry | | | | | | | | | × | | | | | |
| Fagus pacifica Chaney | | × | | | | | | | × | | | | | |
| Fern, fragment | × | | | | | | | | | | | | | |
| Ficus? washingtonensis Knowlton | × | | | | | | | | | | | | | |
| Fraxinus? fruits | | | | | | | | | × | | | | | |
| Ginkgo adiantoides (Unger) Heer | × | | | | | | | | | | | | | |
| Glyptostrobus europaeus (Brongiart) Heer | × | | × | | | | | | × | | | | | |
| Grass, leaf | × | | | | | | | | | | | | | |
| Grass?, genus? | × | | | | | | | | | | | | | |
| Hibiscus occidentalis Berry | | | | | | | | | | | | | | × |
| Hydrangea bendirei (Ward) Knowlton | × | | | | | | | | | | | | | |
| Laurus californica Lesquereux | × | | | | | | | | | | | | | |

TABLE I—*Continued*

| | Spokane Localities | Idaho Localities | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | 1 | 4 | 5 | 7 | 8 | 9 | 12 | 14 | 16 | 17 | 24 | 26 | 27 |
| Laurus grandis Lesquereux | × | | | | | | | | × | | | | | |
| Laurus princeps Heer | × | | | | | | × | | × | | | | | × |
| Laurus similis Knowlton | × | | | × | | | | | × | | | | | × |
| Leaf fungus | | | | | | | | | × | | | | | |
| Leuciscus? bones, spines, and scales (fish) | | | × | | | | | | | | | | | × |
| Libocedrus praedecurrens Knowlton | × | | | | | | | | | | | | | |
| Libocedrus? sp | | | | | | | | | × | | | | | |
| Liquidambar californica Lesquereux | | | | | | | | | × | | | | | |
| Liquidambar pachyphyllum Knowlton | × | | | | | | | | | × | × | | | |
| Liquidambar fruits | × | | | | | | | | × | | | | | × |
| Lycopodium hesperium Knowlton | × | | | | | | | | | | | | | |
| Magnolia dayana Cockerell | × | | | | | | × | | | | | | | × |
| Magnolia sp | × | | | | | | | | | | × | × | | |
| Malva? hespera Knowlton | | | | | | | × | | | | | | | |
| Meibomites lucens Knowlton | × | | | | | | | | | | | | | |
| Moss, genus and species? | × | | | | | | | | | | | | | |
| Myrica lanceolata Knowlton | | | × | | | | | | | | | | | |
| Nyssa knowltoni Berry? | | | | | | | | | × | | | | | |
| Nyssa, small fruits | | | | | | | | | × | | | | | |
| Ostrya oregoniana Chaney? | | | | | | | | | × | | | | | |
| Paliurus knowltoni Berry | | | | × | | | | | | | | | | |
| Phyllites amplexicaulis Knowlton | × | | | | | | | | | | | | | |
| Phyllites crustacea Knowlton | × | | | | | | | | | | | | | |
| Phyllites pardeei Knowlton | × | | | | | | | | | | | | | |
| Phyllites peculiaris Knowlton | × | | | | | | | | | | | | | |
| Phyllites relatus Knowlton | × | | | | | | | | | | | | | |
| Phyllites sophoroides | × | | | | | | | | | | | | | |
| Phyllites sp | × | | | | | | | | | | | | | |
| Pinus monticolensis Berry | | | | | | | | | × | | | | | |
| Pinus sp | × | | | | | | | | | | | | | |
| Platanus dissecta Lesquereux | | | | | | | | | × | | | | | |
| Polytrichites washingtonensis E. G. Britton | × | | | | | | | | | | | | | |
| Populus eotremuloides Knowlton | | | | | | | | | | | | | × | |
| Populus fairii Knowlton | × | | | | | | | | | × | × | × | | |
| Populus heteromorpha Knowlton | × | | | × | | × | × | | × | × | × | | × | × |
| Populus lindgreni Knowlton | × | | | | | | × | | × | × | × | × | | × |
| Populus washingtonensis Knowlton | | | | | | | | | | | | | | |
| Populus sp | | × | × | | | | | × | | | | | | |
| Potamogeton heterophylloides Berry | | | × | | | | | | | | | | | |
| Potamogeton sp | × | | | | | | | | | | | | | |
| Prunus rustii Knowlton | × | | | | | | | | | | | | | |
| Pteris sp | | | × | | | | | | | | | | | |
| Quercus bonseri Knowlton | × | | | | | | | | | | | | | |
| Quercus chaneyi Knowlton | × | | | | | | | | × | | | | | |
| Quercus cognatus Knowlton | | | | | | | × | | | | | | | × |
| Quercus duriuscula Knowlton | | × | | | | | | | | | | | | |
| Quercus horniana Lesquereux | | × | | | | | | | | | | | | |
| Quercus idahoensis Knowlton | | | × | | | | | | | | | | | |
| Quercus merriami Knowlton | × | | | | | | | | | | | | | |
| Quercus obtusa Knowlton | × | | | | | | | | | | | | | |
| Quercus payettensis Knowlton | × | | × | | | | × | | | | | | | × |
| Quercus praenigra Knowlton | × | | | | | | × | | | | | | | × |

TABLE I—*Continued*

| | Spokane Localities | Idaho Localities | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | 1 | 4 | 5 | 7 | 8 | 9 | 12 | 14 | 16 | 17 | 24 | 26 | 27 |
| Quercus cf. Q. pseudolyrata Lesquereux | × | | | | | | | | | | | | | |
| Quercus rustii Knowlton | × | | | | | | | | | | | | | |
| Quercus simulata Knowlton | × | × | | × | | | | | × | | | | × | × |
| Quercus spokanensis Knowlton | × | | | | | | | | | | | | | |
| Quercus treleasii Berry | | | | | | | | | × | | | | | |
| Quercus, acorn | × | | | | | | | | | | | | | |
| Quercus, cup | × | | | | | | | | | | | | | |
| Rhus merilli Chaney | | | | | | | × | | | | | | | × |
| Rhus typhinoides Lesquereux | × | | | | | | | | | | | | | |
| Robinia? sp | × | | | | | | | | | | | | | |
| Salix bryani Knowlton | × | × | | | | | | | | | | | | |
| Salix dayana Knowlton | × | | | | | | | | | × | × | | | |
| Salix elongata O. Weber | × | | | | | | | | | | | | | |
| Salix florissanti Knowlton and Cockerell | | | × | | | | | | | | | | | |
| Salix inquirenda Knowlton | × | | | | | | × | | | | | | | × |
| Salix perplexa Knowlton | × | | | | | | | | | | | | | |
| Salix remotidens Knowlton | × | | | | | | | | | | | | | |
| Salix sp | × | | | | | | | | | | | | | |
| Sequoia langsdorfii (Brongniart) Heer | × | | × | × | | | × | | × | × | × | × | | × |
| Sequoia? sp | × | | | | | | | × | | | | | | |
| Smilax lamarensis Knowlton | | | | | | | | | × | | | | | |
| Sophora alexanderi Knowlton | × | | | | | | | | × | | | | | |
| Sophora spokanensis Knowlton | × | | | | | | | | × | | | | | × |
| Taxodium cone scale | | | | | | | | | × | | | | | × |
| Taxodium dubium (Sternberg) Heer | × | | | | | | | | × | × | × | | | × |
| Taxodium, male aments | × | | | | | | | | | | | | | |
| Taxodium seed | | | | | | | | | × | | | | | |
| Taxodium sp | | | | | | | | × | | | | | | |
| Ternstroemites idahoensis (Knowlton) Berry | | | | | | | × | | | | | | | × |
| Tilia hesperia Berry? | | | | | | | | | | | | | × | |
| Tumion bonseri Knowlton | × | | | | | | | | | | | | | |
| Typha lesquereux Cockerell | | | | | | | | | | | | | × | |
| Typha? sp | × | | | | | | | | | | | | | |
| Ulmus brownelli Lesquereux | | | | | | | | | | | | | | × |
| Ulmus fernquisti Knowlton | × | | | | | | | | | | | | | |
| Ulmus speciosa Newberry | × | | | | | | | | | × | | × | | |
| Umbellularia lanceolata Berry | | | | | | | | | × | | | | | |
| Vaccinium salicoides Knowlton | × | | | | | | | | | | | | | |

## GENERAL CONSIDERATIONS OF THE LATAH FORMATION

The fossil-bearing beds which are described as overlying the pre-basalt terrain in the Spokane–Coeur d'Alene area described by Pardee and Bryan,[1] are found interstratified with the basalt in more than 40 studied localities in Idaho. They are everywhere composed of soft, easily eroded materials such as shale, ash, clay, sand, and

[1] *Op. cit.*

gravel. The material in the Idaho exposures differs little, if at all from that in Spokane area in color and lithologic characteristics, except that at no place was it observed to be arkosic.

### DEFINITION

Although defined in the type locality as being essentially a prebasalt formation, the Latah, as shown by a large number of localities over a much greater area, appears to be more commonly a series interbedded with the lava flows of the Columbia River basalt. In nearly all Idaho localities the greater part of the series has basalt underlying, as well as overlying it. In some places, however, the Latah beds lie directly on the prebasalt terrain. At all places where this was observed, except at Moscow, higher and thicker members interstratified with the overlying basalt were also present.

The beds consist chiefly of fine sediments which contain plant fossils that indicate their Latah age. Where two members of the series are thought to be separated by as much as 400 feet of lava the fossil assemblage shows neither notable break nor evolution, so far as observed. The age of the fossils is generally conceded by paleobotanists to be middle or lower Miocene.[1] Carbonaceous and lignitic material and mineral charcoal also are commonly found.

The various members of the Latah series as found in Idaho appear to occur within a range of 800 feet. The geographic extent of the series is many times greater than was at first surmised.

### STRUCTURE

In the greater part of the area, the Latah beds and the Columbia River basalt flows are nearly horizontal. Their dip of 2 to 4 degrees is chiefly to the west, southwest, or northwest, away from the mountains. But dips as high as 60 degrees are not uncommon in either the lake beds or the lava layers, in the great Uniontown monocline[2] and Craig Mountain anticline, or in the Lewiston-Clarkston downwarp which lies between these folds.

Along Clearwater River and its tributaries the Latah series is exposed in countless landslide blocks which by their nature may be

[1] Pardee and Bryan, *op. cit.*; Knowlton, *op. cit.*

[2] Russell, *op. cit.*

inclined at any angle from the horizontal. The lake beds are interstratified with the basalt without angular unconformity, but both basalt and sediments abut unconformably against prebasalt formations.

## ORIGIN

Field evidence in the Idaho localities persuaded the writers to stand by the time-honored theory of origin which was rejected by Pardee and Bryan[1] for the Spokane area. The original theory, so far as is known, was advanced by Russell[2] and reiterated by Lindgren[3] after extensive reconnaissance by each investigator, not only of the Spokane localities, but also of many of those in Idaho. Restated by the writers, it is essentially as follows:

Westward-flowing drainage in mature valleys would be obstructed and ponded by lava flows advancing toward the mountains from a generally westerly direction along an irregular north-south front. Sediments brought down from adjacent mountains would be deposited in front of the flows at first, and as the ephemeral lakes rose in level and were filled up by deposition, the lakes would extend gradually to the west. The sediments would also thin to the west. Successive lava flows would result in a succession of similar phenomena, and consequently more than one sedimentary series would occur in the section. The final result should be a succession of lava flows alternating with sedimentary beds which lie partly in front of them, and partly interbedded with them. Thickness and area of the part which lay in front of the lava flows would vary greatly from place to place, but the thickness and extent of that part which became interbedded would be more likely to show some uniformity. At all of the Idaho localities, where the relationships are clear, the evidence bears out the postulate. Figure 2, page 9, of the Pardee and Bryan[4] article shows a greater thickness of sediments in front of the early flows and an interleaving with the later basalt flows, with the sediments thinning to the west.

In Figure 3, page 9, of the same article, if the lavas named "valley basalt flows" were assumed to extend beneath the lavas

[1] *Op. cit.*
[2] *Op. cit.*
[3] *Op. cit.*
[4] *Op. cit.*

called "rim rock basalt flows," two sections of Latah would result. The upper layer, overlain by from 300 to 500 feet of basalt, would be about 300 feet thick and would be underlain by the flows now called "valley basalt flows." Beneath the latter, as shown by the Davenport Hotel well, lies a Latah series which by the new assumption would be a lower series. Since that well starts below the top of the "valley basalt flows" it would not show the upper series if these flows actually do continue beneath the "rim rock basalt flows." Should the basalt "sill" shown in their Figure 2 be interpreted as a flow, an upper and a lower sedimentary layer is necessary. This apparently radical assumption for this locality produces two lake-bed layers with lava between them, which is the prevailing condition exposed in the greater number of Idaho localities.

### THICKNESS

The total thickness ascribed to the Latah in the Spokane area[1] is from 1,400 to 1,500 feet. This is a composite section, however, and the greatest measured thickness at any one locality is about 250 feet. The greatest measured thickness of any one member in the Idaho areas is about 350 feet. At no Idaho locality is the combined thickness of the various sedimentary layers believed to be more than 400 feet. The usual thickness of the upper layer is about 100 feet, and that of the lower layer from 150 to 200 feet.

### EXTENT

Lindgren[2] says, regarding extent,

> These sedimentary, intercalated beds can be traced over a large extent of country in the northern and western portions of Nez Perce County and in adjacent portions of Washington and Oregon. They probably extend to the foothills of the Bitterroot Mountains, but as there are other similar layers in that region they have not been identified east of Craig Mountain [p. 76].
>
> Along the canyon of the Clearwater River from Stuart to Peck there are indications of a thick sheet of sandstone and allied rocks at an elevation of about 800 feet below the general surface of the plateau. The true relations are often obscured by landslides. The same, or a similar complex, 200–300 feet thick, is shown throughout the length of Orofino Creek from its mouth to near

[1] Pardee and Bryan, *op. cit.*

[2] *Op. cit.*

Pierce along the walls of the canyon. These beds begin about 750 feet below the surface of the plateau, as along the Clearwater Canyon landslides are very common. Again, sandstone and carbonaceous slate occur at the same level on Little Canyon Creek, draining a part of the broad plateau south of Orofino. Finally, on Potlatch Creek, which joins the Clearwater from the north about 18 miles east of Lewiston, a moderate thickness of sedimentary beds and volcanic dust appears in the basaltic series, 3 miles above the mouth of the canyon. "The Uniontown Plateau is underlain along its southern border and apparently throughout practically its entire area, by horizontal sheets of basalt, which include at least one important bed of clay, sand and gravel, at a depth of about 160 feet below the surface" [p. 76].

Russell,[1] in an article which antedates Lindgren's, says,

The sedimentary beds penetrated in drilling Small's well, and which are exposed in the neighboring gulch, can be traced over a large extent of country in the northern and western portions of Nez Perce County and adjacent portions of Washington and Oregon. They probably extend eastward to the foothills of the Bitterroot Mountains, but as there are other similar layers in that region they have not been identified east of Craig Mountain. All along the sides of the canyon of Snake River from near Lewiston-Clarkston to Buffalo Creek on the east and to the breaks of Grande Ronde Canyon on the west, as well as on the sides of the canyon of the various branches of Asotin Creek, and of Tammany, Sweetwater, and Lapwai creeks, the sedimentary beds referred to are plainly revealed in the topography of the canyon walls [p. 36].

The writers believe the extent of these beds to be essentially limited to the area covered at present or originally by the Columbia River basalt. These beds may be expected to occur at many places along the contact of the lavas with the older rocks which make up the mountainous highlands. Where lava embayments now appear on the maps there were once great valleys of high relief which were obstructed by lava flows and became the sites of temporary lakes of varying sizes, depths, and duration, separated by mountain ranges or low hills.

So far as is known, the northernmost Idaho locality yielding identifiable Latah plants is 2 miles north of Coeur d'Alene. The southernmost Idaho locality yielding a similar flora is 1 mile east of Whitebird. These areas are 140 miles distant from each other by air line. The distance from Whitebird to the exposure near Elk, Washington, is 170 miles by air line. The greatest east and west

[1] *Op. cit.*

range is about 75 miles, as shown by the exposures in Asotin Canyon, Washington, and on the head waters of Orofino Creek.

Latah sediments are exposed at the head of Grand Coulee where, according to Mr. Frank W. McCann,[1] "there seems to be interstratification with the basalt." Fossils collected at this locality were identified by Dr. E. W. Berry of Johns Hopkins University, for the United States Geological Survey. Of 55 determinable species all but 6 are common to the Latah formation. The discovery of this locality greatly extends the area in which the Latah beds occur. The Grand Coulee locality lies 80 miles approximately west of Spokane and about 200 miles northwest of Whitebird. The Latah sediments are now known to be exposed along the margin of the lava area over an irregular front, which has a length of 250 miles.

## LATAH FLORA OCCURRING AT THE SPOKANE AND IDAHO LOCALITIES

Table I, giving the Latah flora, includes the list given for Spokane by Knowlton,[2] and for 13 selected Idaho localities, the numbers of which correspond to similar numbers in the text and on the accompanying map.

Plants from Idaho localities, Stations 1, 4, 5, 7, 8, 9, 12, 14, 26, and 27, were identified by Dr. E. W. Berry of Johns Hopkins University through the courtesy of the United States Geological Survey. Specimens from Stations 16, 17, and 24 were identified by Mr. Erling Dorf of Princeton University, through the courtesy of Dr. Ralph W. Chaney of the Carnegie Institution of Washington.

The writers are also indebted to the United States Geological Survey and to the Idaho Bureau of Mines and Geology for some of the fossil determinations and for other assistance.

In discussing Stations 4 and 5 Dr. Berry says, in connection with his determination, "All but one of the species determined are recorded from the Latah formation of eastern Washington, and that one (*Myrica lanceolata*) is known from only the Payette formation of Idaho. . . . . There is no question of the age of the above plants differing from that of the Latah."

[1] Coulee City, Washington; written communication.

[2] *Op. cit.*

In discussing Stations 8, 9, 14, and 27, he says, "All of these except the new species of *Cercis* and the two marked Bridge Creek are present in the Latah formation around Spokane, Washington." The Bridge Creek species mentioned are *Fagus pacifica* Chaney, and *Ostrya oregoniana* Chaney.

Mr. Dorf said[1] there was no doubt that the specimens from Stations 16, 17, and 24 represent Latah species.

[1] Oral communication.